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REPORT OF INVESTIGATIONS—NO. 176

USE OF THE REFRACTION SEISMIC METHOD FOR DIFFERENTIATING PLEISTOCENE DEPOSITS IN THE ARCOLA AND TUSCOLA QUADRANGLES, ILLINOIS

BY

ROBERT B. JOHNSON

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(*In pocket*)

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USE OF THE REFRACTION SEISMIC METHOD FOR DIFFERENTIATING PLEISTOCENE DEPOSITS IN THE ARCOLA AND TUSCOLA QUADRANGLES, ILLINOIS*

BY

ROBERT B. JOHNSON

ABSTRACT

The purpose of this study is to illustrate the usefulness of the refraction seismic method in distinguishing stratigraphic units within unconsolidated sediments of Pleistocene age. The physical properties of glacial drift are such that compaction can produce considerable changes in seismic velocity. The result of compaction of drift by successive ice sheets in an area should be a velocity stratification which would correspond to stratigraphic units of different ages.

In the area selected for study, the refraction seismograph successfully recorded the various drift sheets and has permitted the accurate mapping of the contact between drifts of Wisconsin and Illinoian ages and further subdivision of Wisconsin drift into Shelbyville and Cerro Gordo age sediments deposited during the Tazewell substage. In the Arcola-Tuscola area, glacial drifts of different ages transmit energy at characteristic average velocities. Reliable subsurface control is essential to successful seismic interpretation of Pleistocene deposits.

CHAPTER I — INTRODUCTION

THE REFRACTION seismograph is an established geophysical tool which is used to obtain depths to bedrock through relatively shallow thicknesses of unconsolidated sediments. It is capable of recording energy which has been produced by a dynamite explosion and propagated through the earth as elastic waves. The elastic waves travel through earth materials at velocities which are partially determined by the elastic properties of the materials. By means of equations based on optical refraction laws it is possible to calculate the depths to many of the levels at which there are velocity changes.

The lowest velocities are recorded from unconsolidated sediments, and the highest usually from igneous and metamorphic rocks. The intermediate velocities found in consolidated sedimentary rocks often are similar to those of the other groups.[1] The principal causes of varying velocities are the interrelated factors of lithology, geologic age, depth of burial, composition, and degree of compaction. The seismic velocity in an unconsolidated sediment depends primarily upon composition, depth of burial, and degree of compaction. A near-surface glacial till may have been compacted to such a degree by an overlying ice sheet that the velocity in it could equal that in a deeply buried deposit of similar composition. The initial porosity of a sediment controls the rate at which compaction will increase the seismic velocity.

The physical properties of glacial drift are such that compaction can produce considerable changes in seismic velocity. If successive ice sheets covered an area, the older sediments and the drift deposited by each ice sheet would be compacted by the weight of each successive ice sheet. The result would be a velocity stratification which should correspond to glacial drifts of different ages, with higher seismic velocities in the older, more compact drifts.

*A dissertation on this subject was submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Geology, Graduate College, University of Illinois, 1954.

[1]Birch, F., et al., Handbook of physical constants: Geol. Soc. Am. Spec. Paper 36, pp. 95-100, 1942.

It was proposed that the refraction seismic method, because it determines the depths at which changes in velocity occur, might be used to differentiate the ages of the sediments resulting from multiple glaciation. An area within the Arcola and Tuscola quadrangles in Champaign, Douglas, and Coles counties was selected for study because: 1) it has a relatively simple Pleistocene history; 2) there is adequate control; 3) there are published reports on geologic interpretation of the Pleistocene stratigraphy; and 4) the level terrain is ideal for refraction seismic operations.

A Century Portable Shallow Refraction Seismograph was used for the research. Whenever possible, seismic stations were located near wells which provided subsurface control. The use of dynamite as an energy source limited the placement of seismic stations because of the highly developed agriculture of the area.

The accuracy necessary for depth determinations was obtained by the exact arrangement of shot energy receivers, or geophones, on the ground surface and by precise interpretation of shot records. Depths were calculated by standard methods.

The most prominent stratigraphic break in the area is between Wisconsin and Illinoian glacial sediments.[2,3] The Wisconsin drift is the result of glaciation during the Tazewell glacial substage,[4,5] when the Shelbyville and Cerro Gordo moraines were formed. The Illinoian drift is undifferentiated in this area. Occurrences of Kansan tills have been reported underlying the Illinoian deposits.[6]

The research conducted in the area reveals that Pleistocene deposits of different ages have representative average seismic velocities. The calculated depths to the velocity changes correspond to the stratigraphic breaks in a sufficient number of instances to illustrate the ability of the refraction seismograph to differentiate accurately the drifts of different ages. The depth to the top of the Illinoian deposits was correctly determined most consistently because of the uniformity of the contact throughout the area. Aside from the few erroneous results expected in any method of geophysical exploration, the occasional failure of the seismograph to indicate drift surfaces can be explained by local geologic conditions which affect the reception of elastic waves.

The seismic data employed in the Pleistocene research were used also to compute depths to the top of bedrock. A comparison is made of the seismic depth determinations with Leland Horberg's bedrock topography map of the area.[7]

Seismic records from the research area frequently depart from those normally expected in refraction work. Previous references to such anomalies are discussed, and a new method for calculation of depths from the abnormal records is proposed. Evidence is presented which shows that changes in the curves are caused by deposits of sand and gravel which are potential aquifers.

It is apparent that, if subsurface control is available as a basis for correlation, the refraction seismic method can provide a means of mapping many Pleistocene deposits resulting from multiple glaciation. The characteristic range of seismic velocities for each stratigraphic unit in any one area offers possible solutions to such problems as age relationships of glacial tills and the areal extent of glacial deposits. The successful application of the seismograph to these problems depends upon a thorough understanding of the factors which control seismic velocities and of the variable geologic conditions which are found in unconsolidated sediments of Pleistocene age.

[2]Horberg, Leland, Pleistocene deposits below the Wisconsin drift in northeastern Illinois: Illinois Geol. Survey Rept. Inv. 165, pp. 25-38, 1953.
[3]Sample studies, Illinois Geol. Survey files.
[4]Leverett, Frank, The Illinois glacial lobe: U. S. Geol. Survey Mon. 38, pp. 191-222, 1899.
[5]Leighton, M. M., The naming of the subdivisions of the Wisconsin glacial age: Science, vol. 77, p. 168, 1933.
[6]Horberg, Leland, *op. cit.*, p. 18.
[7]Horberg, Leland, Bedrock topography of Illinois: Illinois Geol. Survey Bull. 73, pl. 1, 1950.

CHAPTER II — SEISMIC THEORY

Elastic Constants

Seismic exploration is possible because the earth is composed of materials which are elastic. Elasticity may be defined as the "property of a substance to return to its original size and shape after a deforming force has been removed." The deforming force is spoken of as "stress" and the change in size or shape as "strain." In 1678, Robert Hooke stated the principle of elasticity for spring deformation. The principle has since been applied to elastic materials in general and is known as Hooke's Law. A generalized statement of the law is that stress is proportional to strain.

The energy from a dynamite explosion is transmitted, or propagated, through the earth by elastic waves advancing outward from the explosion. The elastic properties of the medium through which the wave is advancing may be described by a number of coefficients of elasticity, or "elastic constants," which are variations of Hooke's Law. Density is not an elastic constant, but it affects the transmission of elastic waves in a material.[1] The velocity of a longitudinal wave is inversely proportional to density and proportional to several elastic constants.

Elastic Waves

Rocks and most of the unconsolidated materials of the earth are elastic and resist an applied stress by a force approximately proportional to displacement. As a result, they oscillate with simple harmonic motion when stress is applied. When some particle of material is displaced, adjacent particles are displaced. The result of increase in number of such displaced particles is the advance of an elastic wave through the material.[2]

Two main categories of waves are found in seismic work: body waves and surface waves. Body waves are so named because they travel through the interior of an elastic medium. Surface waves exist only on the surface of an elastic medium.

Body waves are composed of longitudinal, or primary, waves and transverse, or secondary, waves. They are also known as compressional and shear waves, respectively. A longitudinal wave is so-called because the particles in the wave path move back and forth along the direction in which the wave is traveling. It therefore advances by alternating compressions and rarefactions and is the type that carries the energy which the seismograph records. The name "primary" is derived from the fact that it is the fastest wave and arrives at any given point first from an energy source such as an earthquake focus or dynamite blast. In a transverse wave, the particles in the wave path are displaced along a plane transverse, or at right angles, to the direction of travel. It is known as a "shear" wave, because it advances by means of shearing displacements of the medium, and a "secondary" wave, because it arrives after the longitudinal wave.

Propagation of Elastic Waves

When a column of dynamite is exploded to provide energy for seismic work, the energy released is transmitted or propagated away from the source or shot point in the

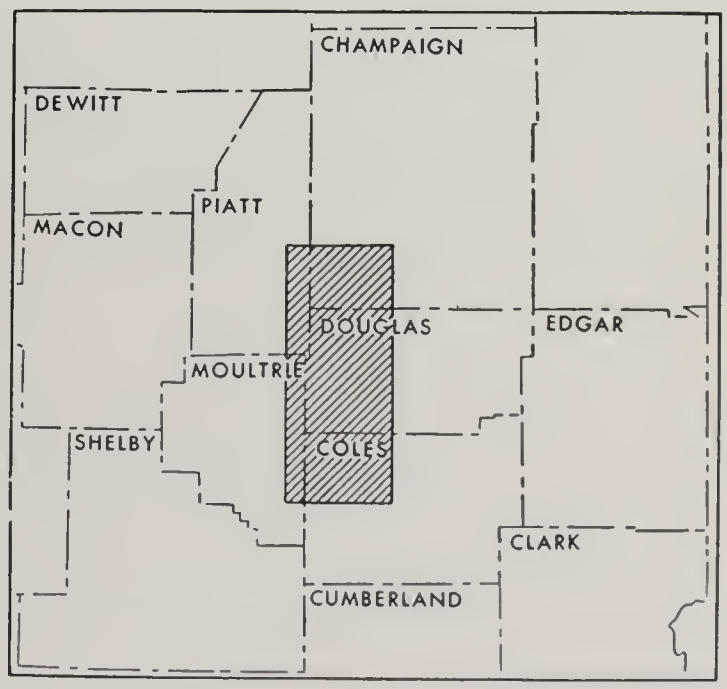


Fig. 1.—Index map of area covered by report.

[1]Leet, L. D., Earth waves: Cambridge, Mass., Harvard Univ. Press, p. 40, 1950.
[2]Leet, L. D., *op. cit.*, pp. 40-41, 1950.

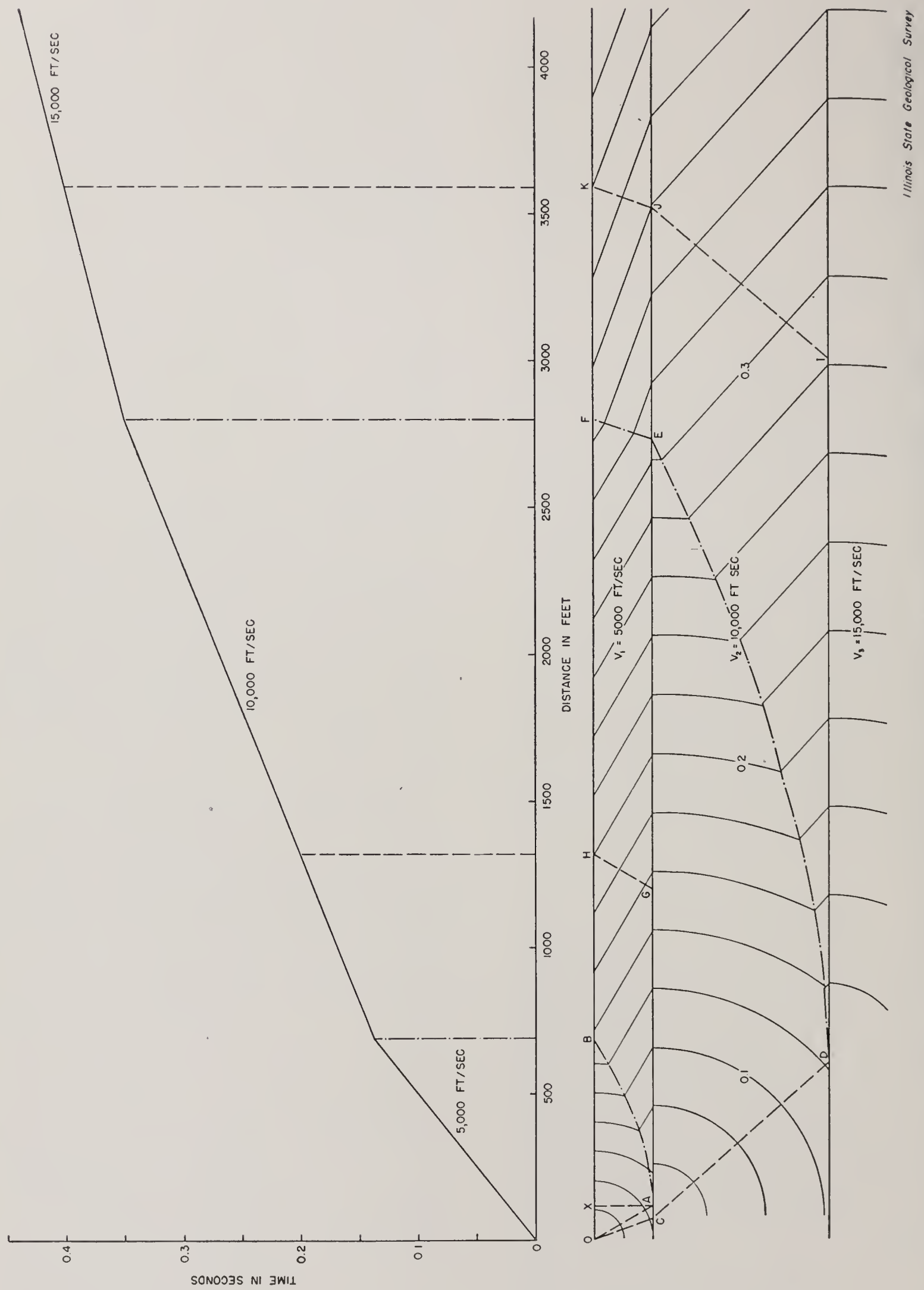


FIG. 2.—Wave fronts, ray paths, and travel-time graph for two horizontal discontinuities.

form of longitudinal and transverse elastic waves. The waves form a family of closed wave fronts that approach more and more closely a spherical shape with increasing distance from the source.[3,4] The wave fronts advance through the medium at a velocity dependent upon the elastic constants of the medium. A wave front is defined as "the surface passing through the most advanced portion reached by a specific disturbance at any particular time."

The shape of the advancing wave front conforms to Huygen's principle. In 1600, Christian Huygen stated that every point on the surface of an advancing wave front is the source of a new wave or wavelet. In a homogeneous and isotropic medium, the wavelets will form an envelope or wave front concentric with the original wave front after some selected time has elapsed. This forms the basis of the construction of a wave-front diagram, a composite figure showing the intersections of various wave fronts with a specific plane at successively equal time intervals.

When an advancing wave front in one medium encounters a boundary with another medium in which the wave front advances with a different velocity, three things occur. Waves enter the second medium as refracted waves, others are reflected back into the first medium, and all generate a reflected and refracted wave of the opposite type. Therefore, if a longitudinal wave were to reach the boundary, the energy in it would be divided into a reflected longitudinal wave, a reflected transverse wave, a refracted longitudinal wave, and a refracted transverse wave.

Figure 2[5] shows three layers in which the velocity of the transmission of energy differs from layer to layer. The diagram shows the fastest wave fronts formed by longitudinal waves, whose ray path in the two upper layers is refracted at the velocity interface at such an angle that the refracted angle is 90°. This angle is known as the critical angle. To the right of point A in the diagram, which is the point where the critical angle occurs, the boundary particles are set in motion by the wave traveling along the top of the lower layer. A wave is generated in the upper medium by this wave and advances into the upper layer. The wave has a straight wave front and its generation is shown in figure 3.[6] Schmidt[7] called this wave the "head wave." If the velocity in the underlying layer were lower than that in the surface layer, the advancing

[3]Heelan, P. A., On the theory of head waves: Geophysics, vol. 18, no. 4, p. 890, 1953.
[4]Ricker, N., Attenuation and amplitudes of seismic waves: Am. Geophys. Union Trans., vol. 30, no. 2, p. 184, 1949.
[5]Leet, L. D., *op. cit.*, p. 117, 1938.
[6]Heiland, C. A., *op. cit.*, p. 506.
[7]Schmidt, O. von, Zur Theorie der Erdbebenwellen: Die 'Wanderender Reflexion' der Seismik als analogen zur 'Kopfwelle' der Ballistike: Zeitschr. Geophysik, vol. 12, pp. 199-205, 1936.

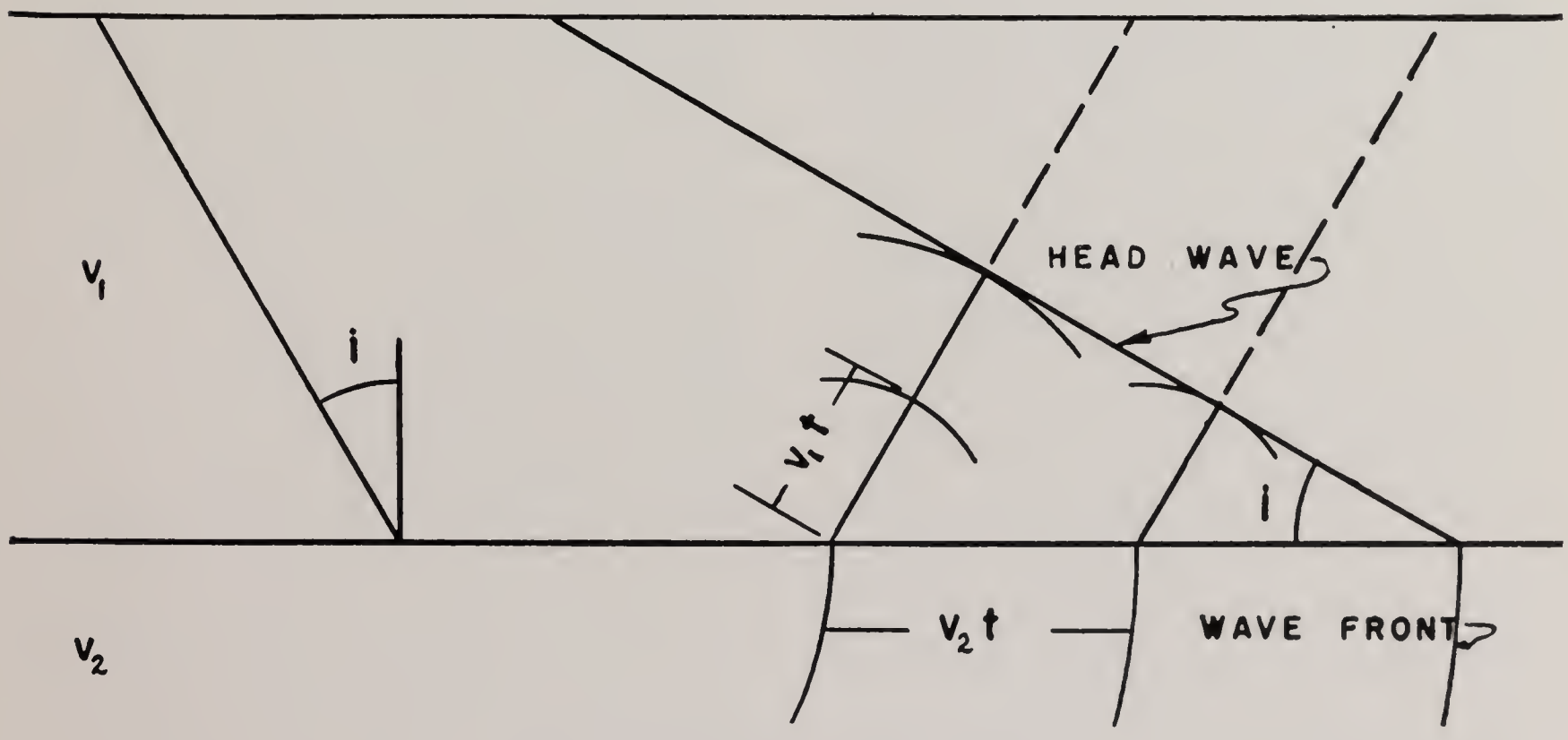


FIG. 3.—Generation of a head wave.

wave front would have been refracted downward, according to Snell's law, and no head wave would have been generated in the upper layer. Snell's law is an optical law, and will be discussed in chapter VII.

The curved line AB in figure 2, designated the coincident-time curve by Thornburgh,[8] is the line along which wave fronts of equal time intersect. Whenever a coincident-time curve reaches the surface there is a break in the travel-time curve, which is plotted in figure 2.

Attenuation of Seismic Energy

In 1941 Born[9] discussed attenuation of seismic energy from viscous and solid-friction losses. For many years prior to 1890 the assumption was that losses were viscous in nature, i.e., they were proportional to the velocity of the wave disturbance. In 1890, Kelvin showed this to be a faulty assumption. It has since been found that the losses in many solid materials are of the type called solid friction. The difference between viscous losses and solid-friction losses may be explained by the stress-strain relations of elastic material under cyclic stress. If the loss is viscous, the energy lost per cycle is proportional to the frequency of the cyclic stress. In solid-friction losses, the energy lost per cycle depends upon the nature of the material and is independent of the frequency.

Born[10] did attenuation experiments with samples of shale, sandstone, and limestone which he assumed to be representative. Tests showed that the losses were from solid friction, and that, in the majority of the materials involved in seismic prospecting, the effect of viscous losses is negligible over the range of frequencies used, that is, below 150 cycles per second (c.p.s.). Although the tests were made at room temperature and atmospheric pressure, it was believed that the effects of temperature and pressure would alter only the magnitude of the attenuation and not its nature. Clewell and Simon[11] agreed that below 150 c.p.s. the viscous losses were unimportant. A question arises concerning the nature of losses in unconsolidated sediment. Born[12] surmised that such deposits would exhibit a plastic behavior and hence produce appreciable viscous losses. However, sediments at some depth might be compacted enough by the overburden to behave like solids and cause solid-friction losses.

Born[13] and Clewell and Simon[14] agree that the earth acts as a low-pass filter, which discriminates against high frequencies increasingly the longer the wave path. This discrimination is often encountered in reflection work when records from deeper horizons show energy concentrated in the frequency band lower than that of reflection from shallower horizons.[15] Records of earthquakes show low frequencies after many hundreds or thousands of miles of travel. Refraction records also show frequencies considerably lower than reflection records show because of the greater distance traveled for any specific depth.[16]

High frequencies are also eliminated by other factors. Since the earth is not homogeneous, stratification and numerous small irregularities in vertical and horizontal directions cause diffusion of seismic wave energy. The intensity of the scattering is a function of the wave length and the size of the irregularity. The greatest amount of scattering is believed to be in the near-surface weathered zone.[17]

[8]Thornburgh, H. R., *op. cit.*, pp. 190-191.
[9]Born, W. T., Attenuation constant of earth materials: Geophysics, vol. 6, no. 2, pp. 132-148, 1941.
[10]Born, W. T., *op. cit.*, pp. 133-141.
[11]Clewell, D. H., and Simon, R. F., *op. cit.*
[12]Born, W. T., *op. cit.*, p. 142.
[13]*Ibid.*, p. 147.
[14]Clewell, D. H., and Simon, R. F., *op. cit.*
[15]Born, W. T., *op. cit.*, p. 147.
[16]Clewell, D. H., and Simon, R. F., *op. cit.*
[17]*Ibid.*, pp. 55-56.

CHAPTER III — PREVIOUS SHALLOW REFRACTION WORK

Prior to use in shallow exploration, the refraction seismograph gained prominence in long-range surveys for buried salt domes in the search for petroleum.[1] As exploration for other types of oil traps increased, the reflection method gradually became the most widely used of the two seismic methods, and references to refraction work in the literature became very few. In the last several years, however, there has been an increase in the number of reports published on refraction work, which indicates the increased use of the seismograph for shallow exploration. The refraction method is the better suited for such work, although a recent paper[2] indicates that new shallow techniques are being developed for the reflection seismograph. The bulk of the information published on shallow refraction work has dealt with engineering problems and bedrock depth determinations. Various lithologic units within sediments of Pleistocene age have been recognized, but no attempt has been made to map such units or correlate them with known stratigraphy. It is the purpose of this study to provide information on this phase of refraction exploration.

One of the early applications of the refraction seismograph was described by Edge and Laby in 1931[3] in connection with the tracing of buried channels in Australia in a search for placer gold deposits. Three years later in this county, Partlo and Service[4] mapped near-surface faults, their work being of interest because the blow of a sledge with an ingenious timing device was used for energy instead of dynamite for the shallower work. A paper by Shepard,[5] published in 1935, describing early equipment, shot records, and simple depth calculations, was one of the first articles in which engineering applications such as dam sites and highway construction were discussed. The Corps of Engineers of the United States Army adopted the refraction seismograph as an exploration tool in 1938.[6] Shepard[7] discussed the information gathered during the first year of Army operation in a paper which presented for the first time many of the problems confronted in research on unconsolidated material. Compaction from overburden was believed to play a larger role in the seismic velocity of a material than the character of the original material.

In 1940, Shepard and Wood[6] described several river improvement and dam projects in which the refraction seismic method was invaluable. The principal contributions of the paper were the statements that the water table could usually be located and that it is often difficult to determine the depth to certain types of bedrock when the bedrock is overlain by a well compacted till. Shepard and Haines[8] found that the seismic data were useful in estimating depths to individual strata in the overburden for studies of structure sites. Another discovery was that the soil profiles as indicated by the seismograph were not always recognizable in driller's logs or samples, for velocities are determined by depth of frost action and compaction in addition to composition. In 1947 Coster and Gerrard[9,10] published two papers describing seismic profiles made in England for the determination of the thickness of Pleistocene-age valley fills. The only significant information in either paper came from a remark concerning a "standard

[1]Barton, D. C., The seismic method of mapping geologic structure: Am. Inst. Min. Met. Eng. Trans., Geophysical Prospecting: vol. 81, pp. 572-624, 1929.

[2]Allen, C. F., Lombardi, L. V., and Wells, W. M., The application of the reflection seismograph to near-surface exploration: Geophysics, vol. 17, no. 4, pp. 859-866, 1952.

[3]Edge, A. B. B., and Laby, T. H., The principles of geophysical prospecting: Cambridge Univ. Press, 1931.

[4]Partlo, F. L., and Service, J. N., Seismic refraction methods as applied to shallow overburdens: Am. Inst. Min. Met. Eng. Trans., Geophysics, vol. 110, pp. 473-492, 1934.

[5]Shepard, E. R., Subsurface exploration by earth resistivity and seismic methods: Public Roads, vol. 16, no. 4, pp. 57-67, 1935.

[6]Shepard, E. R., and Wood, A. E., Application of seismic refraction method of subsurface exploration to flood control projects: Am. Inst. Min. Met. Eng. Trans., Geophysics, vol. 138, pp. 312-325, 1940.

[7]Shepard, E. R., The seismic method of exploration applied to construction projects: Military Engineer, vol. 31, no. 179, pp. 370-377, 1939.

[8]Shepard, E. R., and Haines, R. M., Seismic subsurface exploration on the St. Lawrence River project: Am. Soc. Civil Eng. Proc., vol. 68, no. 10, pp. 1743-1761, 1942.

[9]Coster, H. P., and Gerrard, J. A. F., A seismic investigation on the outflow of Windermere: Geol. Mag., vol. 84, no. 6, pp. 224-228, 1947.

[10]Coster, H. P., and Gerrard, J. A. F., A seismic investigation of the history of the River Rheidol in Cardiganshire: Geol. Mag., vol. 84, no. 6, pp. 360-368, 1947.

error of approximately ± 10%" in refraction seismic depth determinations.

Applications of the seismograph in road construction, bridge construction, and location of fill in New England were described by Linehan.[11] In the course of this work, the difficulties of a high-velocity till and a lower-velocity bedrock were encountered. Closely spaced, or continuous, profiles were said to be the solution to such a problem.

One of the earliest publications to mention the application of refraction seismology to groundwater exploration was written by Linehan and Keith in 1948.[12] Their work consisted mainly of bedrock determinations, but velocity measurements were made in the unconsolidated sediments. In Connecticut where the work was done, characteristic velocities were noted for fine dry sand, wet sand, wet or dry gravel, and poor aquifers such as hardpan, boulder clay, and old till. Apfel and Linehan[13] wrote of instances in which shallow refraction seismic work was of assistance in highway construction. In 1949, a Swedish paper[14] described bedrock depth determinations and glacial ice thickness studies in a general fashion.

In a paper delivered before the 1950 convention of the American Water Works Association, Bays[15] mentioned the probable expansion of the use of seismic data in groundwater prospecting. Another reference to the use of the seismograph in the exploration for foundation sites for engineering structures was published in 1950[16] in a trade journal. Two papers by Moore[17,18] appeared in print in 1950 briefly describing the operation of the shallow refraction seismograph and the computation of simplified data. Applications of the refraction method to bedrock depth determination were described. The use of electrical earth resistivity and seismic methods in groundwater exploration were discussed by Stickle, Blakely, and Gordon in 1952.[19] General statements were made about water table location and bedrock top determination using the seismic method.

In a paper published in 1952,[20] Wesley described seismic operations in Detroit, in which a sledge hammer blow or a portion of a stick of dynamite was used for energy. The travel-time graph shown in the paper has a delay, or time discontinuity, in one portion of one curve. Test drilling indicated the presence of a relatively thin layer of sand, which was said to be the cause of the delay. This is one of the few references to a travel-time delay and its cause. No information was given for bedrock depth computation for the curve with the delay.

Evison[21] has described a seismic survey made in New Zealand for the depth to a high-velocity bedrock through a thick sand layer. He was dissatisfied with the accuracy, and he found that the best results were obtained from shooting parallel to the bedrock strike. Evison also stated that the refraction technique is inadequate to resolve details of a shallow interface and that it is impossible to resolve two successive interfaces if the distance between them is less than a certain fraction of their depth. The cause of these problems was said to be the nature of the explosive impulse. It was concluded that using dynamite as the energy source seriously limits the use of the refraction method in shallow exploration. An electro-magnetic energy source was proposed, but many problems needed to be solved before it could be used.

Bird[22] has described several engineering applications of the seismograph. He mentioned that geologic discontinuities in com-

[11]Linehan, D., Seismology as a geologic technique: Highway Research Board Bull. 13, pp. 77-85, 1948.
[12]Linehan, D., and Keith S., Seismic reconnaissance for groundwater development: Jour. New England Water Works Assoc., vol. 63, no. 1, pp. 76-95, 1948.
[13]Apfel, E. T., and Linehan, D., Geologic interpretation of seismic data: Highway Res. Bd. Bull. 13, pp. 90-99, 1948.
[14]Hedström, H., and Kollert, R., Seismic sounding of shallow depths: Tellus, vol. 1, no. 4, pp. 24-36, 1949.
[15]Leggette, R. M., Bays, C. A., Thorpe, T. W., Ferris, J. G., and Baffo, J. J., Prospecting for groundwater: Am. Water Works Assoc. Jour., vol. 42, no. 10, pp. 945-965, 1950.
[16]Rose, R. B., Shallow seismic exploration of foundation sites for engineering structures: Southwest Builder and Contractor, June 9, 1950.
[17]Moore, R. W., Development of geophysical methods of subsurface exploration in the field of highway construction: Highway Res. Bd. Bull. 28, pp. 73-99, 1950.
[18]Moore, R. W., Geophysical methods of subsurface exploration in highway construction: Public Roads, vol. 26, no. 3, pp. 49-64, 1950.
[19]Stickle, J. F., Blakely, L. E., and Gordon, B. B., Geophysics and water: Jour. Am. Water Works Assoc., vol. 44, no. 1, pp. 23-35, 1952.
[20]Wesley, R. H., Geophysical exploration in Michigan: Econ. Geology, vol. 47, no. 1, pp. 57-63, 1952.
[21]Evison, F. F., The inadequacy of the standard seismic techniques for shallow surveying: Geophysics, vol. 17, no. 4, pp. 867-875, 1952.
[22]Bird, P. H., Experience with geophysics in New York State: Am. Soc. for Testing Materials Spec. Tech. Pub. 122, pp. 151-155, 1952.

plex glacial deposits have been identified. More applications of the refraction method to engineering problems were discussed by Linehan.[23] He gave information concerning field operations and representative velocities for shallow unconsolidated materials.

In 1953, several seismic profiles were made in England to determine the thickness of unconsolidated deposits overlying shallow mine workings.[24] Problems of delays in the travel-time curves will be discussed in chapter VII.

[23]Linehan, D., Seismology applied to shallow zone research: Am. Soc. for Testing Materials Spec. Tech. Pub. 122, pp. 156-170, 1952.

[24]Brown, P. D., and Robertshaw, J., A seismic survey-determination of the thickness of unconsolidated deposits overlying shallow mine workings: Colliery Guardian, vol. 187, no. 4829, pp. 347-353, 1953.

CHAPTER IV — FACTORS CONTROLLING SEISMIC VELOCITY

Seismic velocity is affected by a number of factors which account for the variability and overlapping of recorded velocities. An adequate understanding of those factors is necessary to minimize the chances for error in interpretation. A few of the more important controlling conditions will be discussed in this chapter.

Depth of Burial

Depth of burial of a material is an important factor influencing the velocity at which energy will travel through a medium. Faust[1] has shown that for any geologic period or epoch, velocities increase proportionally to depth, with limestone exhibiting a smaller increase than sandstones and shales. The velocity in any one formation was also found to increase with depth. Under similar conditions the velocities in sandstones were found to be greater than those in shales. The rate of increase of velocity with depth is greatest at shallow depths,[2] indicating that a factor easily affected by compaction causes the increased rate of velocity change. This factor is the porosity of the material. The weight of a thick ice sheet may cause porosity changes similar to those resulting from deep burial.

West[3] suggests that rock density is directly related to porosity, and thus seismic velocity increases with a decrease in porosity. Elasticity has been shown to increase with density,[4] and therefore it too is dependent upon porosity. Weatherby and Faust[5] showed that lithification must increase the elastic constants at a faster rate than the increase in density to account for the increased seismic velocities found with increasing lithification. It was believed to be a valid assumption that elasticity generally increases more rapidly than density, since density seldom extends beyond a range of 1.3 to 3.0 gm/cm^3, whereas elasticity may increase ten or more times, especially if there has been recrystallization. Krumbein[6] has stated that the depth of overburden or other postdepositional changes might be as important as the original sediment characteristics in determining final elastic properties. Haskell[7] has shown that added compaction does not affect shale as much as sandstone, which emphasizes the importance of porosity.

A thin surface layer with low velocity, known as the weathered layer, has been recognized by reflection seismic workers since about 1930. It is almost universally present and has been differentiated by its low velocity rather than by any significant physical difference between it and the materials directly below it. Lester[8] found that the velocity of sound in this layer was always comparatively low, ranging from 2000 to 2500 ft./sec., whereas the velocities in the same kind of material immediately below it averaged 5600 ft./sec. He indicated that in the Gulf Coast region the layer was not a true weathered layer. In that area the geophysically determined zone was found to be almost uniform and thin, although the geologic weathered zone was deep and variable in thickness. In several instances velocities were recorded that were lower than the velocity of sound in air, which gave rise to the consideration of the fact that air was included in the surface materials. A velocity as low as 550 ft./sec. was obtained. To the zone of low velocity Lester gave the name "aerated surface layer," which seems to be the more correct term to apply to it.

[1]Faust, L. Y., Seismic velocity as a function of depth and geologic time: Geophysics, vol. 16, no. 2, pp. 192-206, 1951.

[2]Hughes, D. S., and Jones, H. J., Elastic wave velocities in sedimentary rocks: Am. Geophys. Union Trans., vol. 31, pp. 169-173, 1950.

[3]West, S. S., Dependence of seismic wave velocity upon depth and lithology: Geophysics, vol. 15, no. 4, p. 657, 1950.

[4]*Ibid.*

[5]Weatherby, B. B., and Faust, L. Y., Influence of geological factors on longitudinal seismic velocities: Am. Assoc. Petr. Geol. Bull., vol. 19, no. 1, pp. 1-8, 1935.

[6]Krumbein, W. C., Some relations among stratigraphy and seismic explorations: Am. Assoc. Petr. Geol. Bull., vol. 35, no. 7, p. 1514, 1951.

[7]Haskell, N. A., The relation between depth, lithology, and wave velocity in Tertiary sandstone and shale: Geophysics, vol. 6, no. 4, pp. 318-326, 1941.

[8]Lester, O. C., Jr., Seismic weathered or aerated layer: Am. Assoc. Petr. Geol. Bull., vol. 16, no. 12, p. 1231, 1932.

Lithology

Some of the effects of lithology on seismic velocities have been referred to above. Kisslinger's paper[9] on the effect of chemical composition on the velocity of seismic waves in the carbonate rocks indicates how important lithology may be even within one class of rocks. The density and elasticity of a rock depend on the constituent materials and the manner of mineral assembly. The manner of mineral assembly has its greatest influence on the physical properties of shales and sandstones. Kisslinger showed that, although limestones are also controlled by these factors, the observed increase of velocity in dolomitic limestone of formation magnitude over the velocity in calcitic limestone of the same age may be explained on the basis of the relations between the elasticities and densities of the minerals calcite and dolomite. The effect of the lime content of sandstone and shale on seismic velocity has been studied by Houston,[10] who found that lime content is directly proportional to the seismic velocity.

Very little quantitative information appears in the literature on seismic velocities in unconsolidated sediments. Shepard[11] states that the origin of a soil appears to have less influence on velocity than age, moisture, texture, compaction, void space, and cementation. The composition of the material is also an important factor. Sand and gravel can usually be differentiated from glacial till by means of velocity measurements. Variations in composition within a deposit such as a till may also affect the velocity, but the extent of this influence may be masked by other factors such as compaction and water content.

The most exhaustive investigation of unconsolidated sediments appears to have been made by Iida.[12] Although the paper was concerned primarily with the effect of water content on the elastic constants and seismic velocity of a material, valuable information about the effect of composition was included. At zero water content the velocity of a longitudinal wave decreases as the silt and clay content increases. There is a reduction from the maximum velocity when the clay fraction predominates. The highest density occurs in a sample containing the most sand. The elastic property measurements increase steadily as the silt and clay percentages increase; decreasing density and increasing elasticity result in increasing velocity.

Water Saturation

Water content is an important factor controlling seismic velocity, especially in unconsolidated sediments. Iida[13] studied the effect of water content on seismic velocity, density, and several elastic constants. Soil density was observed to increase with increased water content until a critical content is reached, after which density decreased with additional water. The critical water content differs with different kinds of soil. Iida concluded that in general soil density increases with increase of clay and decreases with increase of sand.

The velocity of elastic waves tends to decrease with an increase of water until a critical water volume is reached. This critical value is at a higher percentage than for the inflection point in the density curve. The rate of decrease of velocity increases quite rapidly with the increase in clay.

Hughes and Kelly[14] have studied the variation of elastic wave velocity with pressure and water saturation in sandstone. At low pressures in the magnitude of 50 atmospheres, the velocity rises sharply at saturations up to 10 percent. It remains constant from 10 to 90 percent and decreases as it approaches 100 percent. As pressure increases, the rise in velocity at low saturations decreases, and stops at 500 atmospheres. Born[15] found that increased moisture in an

[9]Kisslinger, Carl, The effect of variations in chemical composition on the velocity of seismic waves in carbonate rocks: Geophysics, vol. 18, no. 1, pp. 104-115, 1953.
[10]Houston, C. E., Seismic paths assuming parabolic increase of velocity with depth: Geophysics, vol. 4, no. 4, pp. 242-246, 1939.
[11]Shepard, E. R., The seismic method of exploration applied to construction projects: Military Engineer, vol. 31, no. 179, p. 373, 1939.
[12]Iida, K., On the elastic properties of soil, particularly in relation to its water content: Tokyo Imp. Univ. Earthquake Res. Inst. Bull., vol. 18, no. 4, pp. 675-691, 1940.
[13]*Ibid.*
[14]Hughes, D. S., and Kelly, J. L., Variation of elastic wave velocity with saturation in sandstone: Geophysics, vol. 17, no. 4, pp. 739-752, 1952.
[15]Born, W. T., Attenuation constant of earth materials: Geophysics, vol. 6, no. 2, pp. 135-141, 1941.

oven-dried sandstone sample decreased the velocity of wave propagation.

Many writers agree that there is a sharp increase in velocity when the water table is encountered.[16,17,18,19] White and Sengbush[20] have observed that all velocities in loose near-surface sand are very low at the surface and increase gradually with depth. At the water table an abrupt increase in compressional speed occurs with no perceptible effect on the shear properties of the material.

A comparison of the field evidence with the experimental work by Iida indicates that field evidence has provided results the opposite of those obtained in the laboratory. The contradiction could be due to the fact that in Iida's experiments the samples remained at a constant compaction, which is not maintained in nature. Conditions within the aerated zone in the field could not be duplicated in the laboratory. A valid comparison, therefore, may not be possible between field and experimental results.

[16]Shepard, E. R., and Wood, A. E., Applications of seismic refraction method of subsurface exploration to flood control projects: Am. Inst. Min. Met. Eng. Trans., Geophysics, vol. 138, p. 320, 1940.
[17]White, J. E., and Sengbush, R. L., Velocity measurements in near-surface formations: Geophysics, vol. 18, no. 1, p. 61, 1953.
[18]Linehan, D., and Keith, S., Seismic reconnaissance for groundwater development: Jour. New England Water Works Assoc., vol. 63, no. 1, p. 82, 1948.
[19]Lester, O. C., Jr., *op. cit.*, p. 1233.
[20]White, J. E., and Sengbush, R. L., *op. cit.*

Geologic Structure

It is known that the apparent velocity of seismic energy in a medium for an up-dip shot is greater than the true velocity of the medium below the dipping surface and that it is less in a down-dip shot. In figure 4, the ray paths for an up-dip and a down-dip shot are shown. In the down-dip shot, the energy has to travel through more of the lower velocity, V_1, material than in the up-dip shot, each starting at the same depth. Since velocity is a function of time and distance, in the down-dip instance the greater travel time for the same horizontal distance results in a lower-than-normal velocity. The dipping surface may not be the plane between two dipping rock formations; instead it may be the irregular surface of the bedrock or a velocity discontinuity within overlying unconsolidated sediments. In such cases, the seismic results tend to average the surface. Since a horizontal surface is a rarity in geology, most seismic refraction profiles are shot at both ends. This procedure gives the computer the up- and down-dip apparent velocities, from which more accurate depth determinations may be made.

Another aspect of structural control may be found in areas of very steeply dipping beds. Bird[21] has noted considerable errors

[21]Bird, P. H., Experience with geophysics in New York State: Am. Soc. for Testing Materials Spec. Tech. Pub. 122, p. 153, 1952.

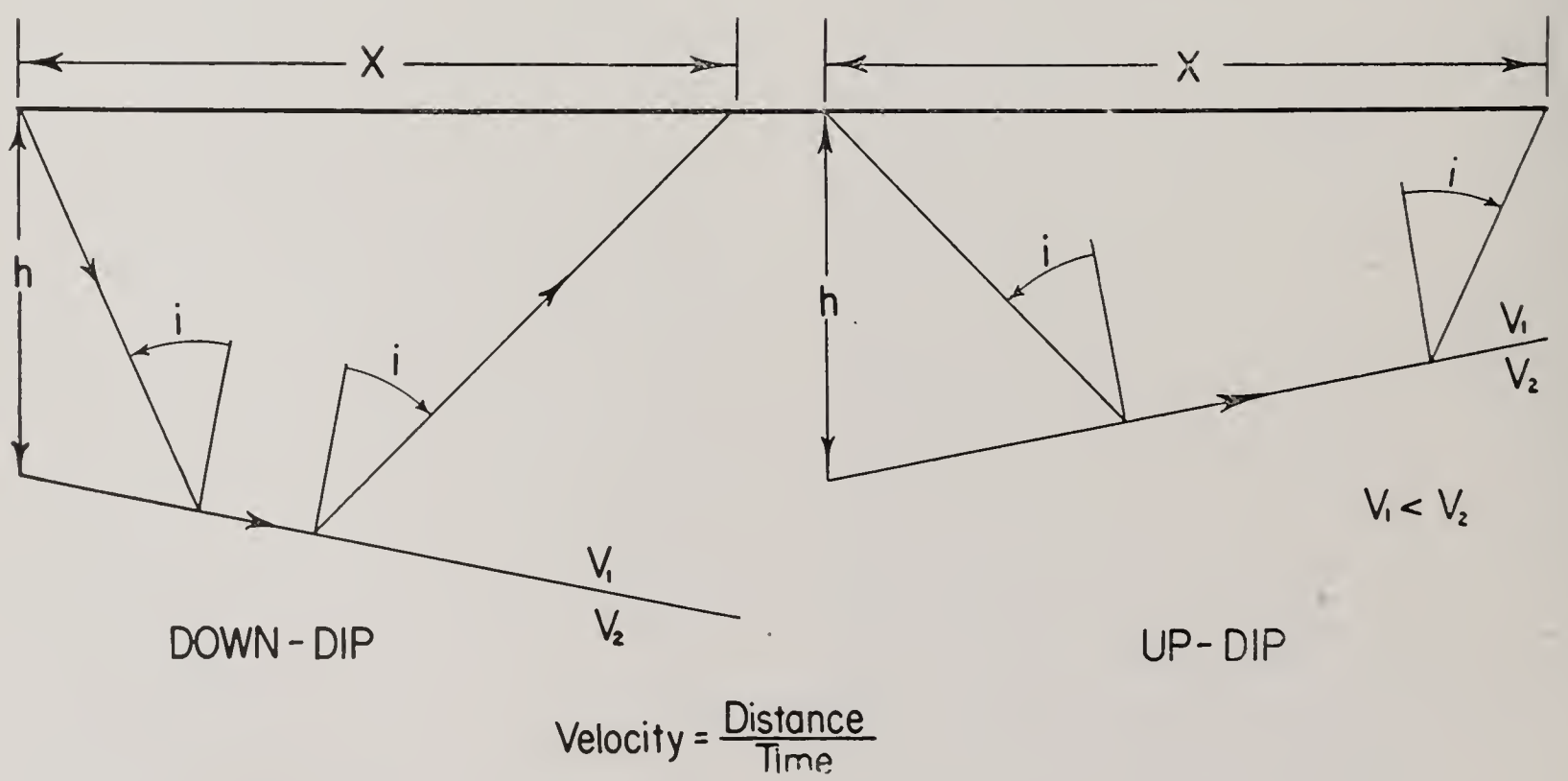


Fig. 4.—Down-dip and up-dip ray paths.

in depth computations when seismic profiles have been made perpendicular to the strike of the beds. Profiles parallel to the strike are more accurate. When shooting across the strike, the ray paths from one end differ considerably from those from the other end of the profile because of the steeply dipping interfaces which propagate some of the energy.

The problem of dipping surfaces was encountered in almost all the profiles made during this study. Standard methods of calculation were used. Steeply dipping beds have been found only in a few places along the west flank of the LaSalle anticline[22] west of Tuscola, Illinois. The accuracy of the depth determinations is not known because of the lack of reliable well data.

Geologic Age

The effect of geologic age on seismic velocity in a sediment is usually closely associated with depth of burial. Faust[23] found that the mean velocities in rocks increase with age, shales and sandstones exhibiting a greater change than limestones.

Experimental Data

Although there are a number of statements in the literature concerning the relationships between seismic velocity and composition and compaction of a material, it seemed advisable to do some experimental work on the subject. The conditions at station 24, SW¼ SE¼ sec. 25, T. 16 N., R. 8 E., Tuscola quadrangle (plate 2), were found to be ideal for such a study. The seismic profile at station 24 was 1750 feet long and was laid east-west. The east end was at an elevation of 679 feet above sea level and the west end at 662 feet, estimated from the topographic map. Immediately underlying the east end were 7 feet of material in which the seismic velocity was 1863 ft./sec., 10 feet of material in which velocity was 3750 ft./sec., and a greater thickness in which the velocity was 5882 ft./sec. Underlying the west end were 7 feet of material of 2240 ft./sec. velocity and a layer of material of 5000 ft./sec. velocity. The topography indicated that the Kaskaskia River has eroded the prairie eastward about one-half mile from the present river bed to form the lower elevation on which the west end of station 24 is located. In this erosion process, the layer of material of 3750 ft./sec. velocity seems to have been eliminated from the western half of the profile. The top 7-foot layer is the aerated layer described by Lester.[24] An explanation of the difference between velocities at the ends is that the velocity-reducing agents operating in the upper zone acted upon glacial tills with different original velocities, namely, 3750 ft./sec. velocity at the east end and 5000-5882 ft./sec. velocity at the west end. The 5000 ft./sec. and 5882 ft./sec. velocity materials are assumed to be the same till, the difference in velocity resulting from the reduction of overburden pressure by erosion of material from the west end of the profile.

Soil samples were taken at each end of station 24 by means of a soil auger, sample A from the 9 to 10 foot depth interval at the east end and sample B from the 8 to 9 foot depth interval at the west end. The samples were studied to ascertain any significant composition variations which might explain the difference in velocity at the ends. The data obtained were plotted as histograms (figs. 5 and 6) which follow the Wentworth[25] classification, in which the sand grade ranges from 1.981 mm through .062 mm, the silt grade from .0442 through .0039 mm, and the clay grade is smaller than .0039 mm.

An examination of the histograms shows the tri-modal character of each sample. The greatest variations occur in the materials smaller than .0078 mm and larger than .00098 mm. If the percentages for the sand, silt, and clay grades are totaled, sample A has 31.76% sand, 57.67% silt, and 4.96% clay, whereas sample B has 31.15%

[22]Weller, J. M., et al., Geologic map of Illinois: Illinois Geol. Survey, 1945.
[23]Faust, L. Y., *op. cit.*
[24]Lester, O. C., Jr., *op. cit.*, pp. 1230-1234.
[25]Wentworth, C. K., A scale of grade and class terms for clastic sediments: Jour. Geology, vol. 30, pp. 377-392, 1922.

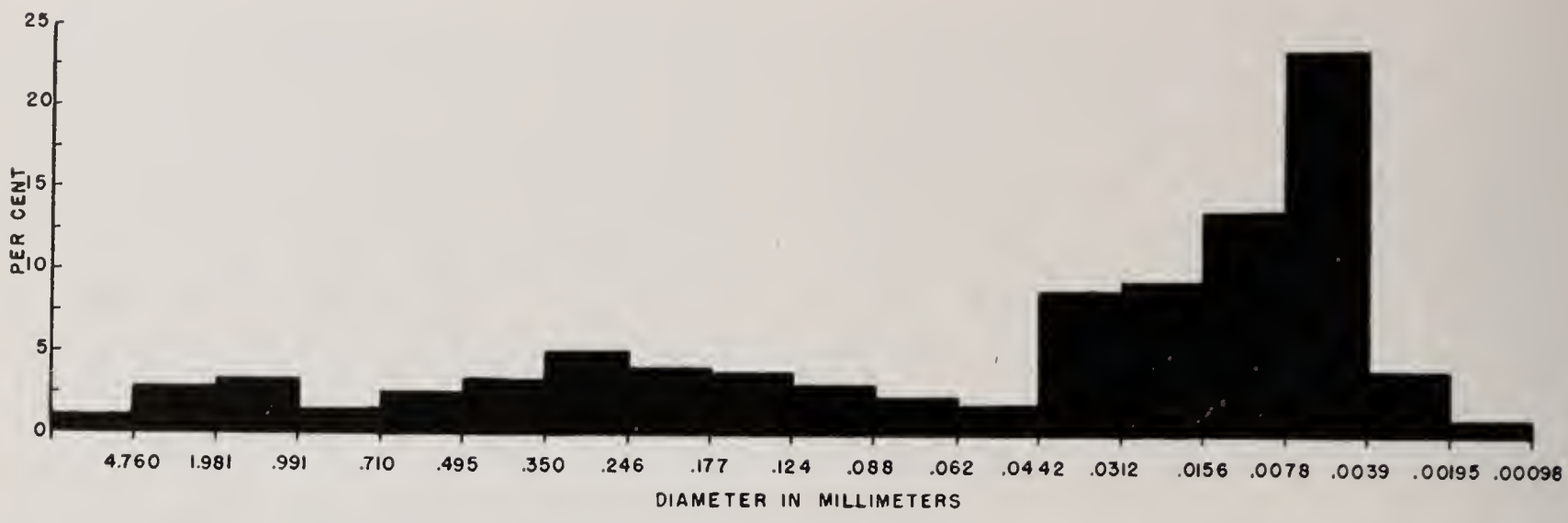


FIG. 5.—Histogram of glacial till sample A, station 24A, sec. 25, T. 16 N., R. 8 E., Tuscola quadrangle.

sand, 45.31% silt, and 17.87% clay. Iida[26] has shown that an increase in clay-size sediments increases the seismic velocity of a material. The reduction in the percentage of sand in Iida's sample seems to exert some appreciable influence on the increase of velocity also. Since the sand percentages in the two samples are almost the same and since the differences in silt-clay percentages are the result of grade-size fluctuations very close to the arbitrary break between silt and clay, it may be assumed that a variation in composition is not the major cause of the velocity differential noted between samples A and B.

Compaction from the weight of the overlying material and the weight of an overriding ice sheet is assumed to be the major cause of the difference. The weight of the overlying material probably has contributed at least 882 ft./sec. of the difference, as this is the difference between end A, where there is an overlying layer, and end B, where this layer has been eroded away. The amount attributable to this cause undoubtedly is greater than the figure given because of the failure of the material to regain its original porosity and fabric after the load was removed. Some fraction of the difference between 3750 ft./sec. and 5000 ft./sec. has then probably been caused by an overlying ice sheet. A thin unconsolidated layer, in which the velocity is as low as 3750 ft./sec., below the low-velocity surface layer is not widespread in the Arcola-Tuscola area.

It is important to keep in mind that compaction of glacial drift can be caused by an overlying sheet of ice as well as overlying sediments. It can be expected that any layer will be compacted to some extent by overlying material. Thus one might expect a gradual increase in velocity with depth in unconsolidated sediments; however, velocity increases are discontinuous, sometimes as much as 2000 to 3000 ft./sec. Throughout the Arcola-Tuscola area, the velocity changes occur in the majority of the instances at elevations which are remarkably consistent. It is proposed that the weight

[26]Iida, K., *op cit*, pp 676-678..

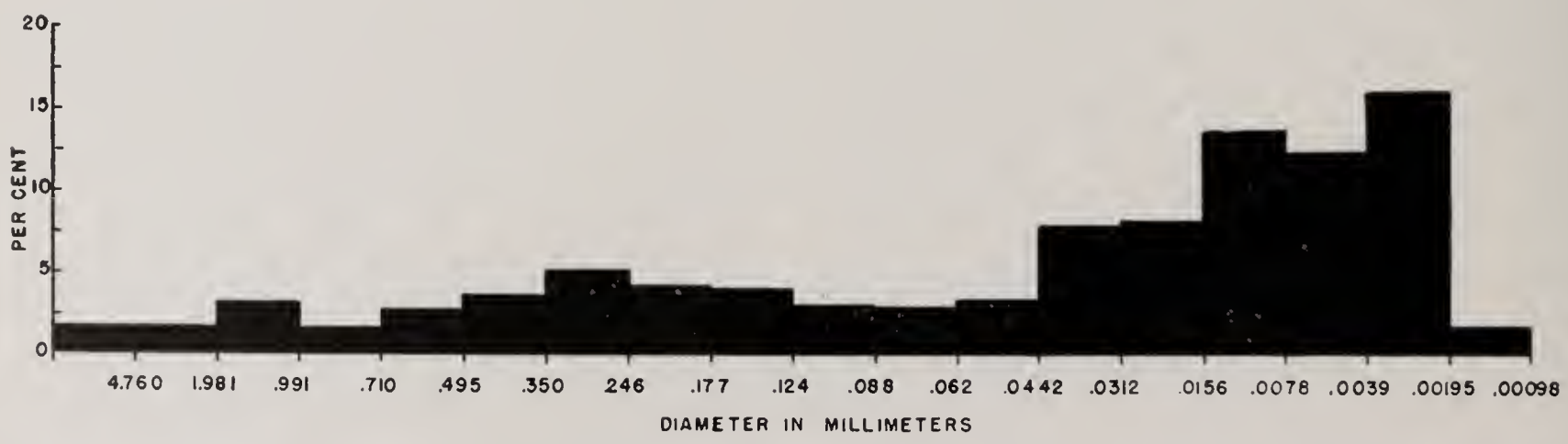


FIG. 6.—Histogram of glacial till sample B, station 24B, sec. 25, T. 16 N., R. 8 E., Tuscola quadrangle.

of an ice sheet over the area would have compressed the older underlying sediments so that velocities in them would be consistently higher than in the material deposited by the overlying ice, either as ground or ablation moraine or as outwash. Horberg[27] noticed that Illinoian till under Wisconsin till was much more compact and jointed than the Wisconsin material. This was attributed to compaction under the weight of Wisconsin ice and to the silty composition of the Illinoian till. MacClintock[28] found that exposures of pre-Illinoian till were dense and hard in comparison with overlying Illinoian till in an area south of the limit of Wisconsin glaciation.

Seismic shots were twice made in water-saturated ground. The seismic velocities shown for the saturated material on the travel-time graphs differ considerably from those from dryer ground nearby. In both instances the unconsolidated material was alluvium in which velocity is low. If the material had been glacial drift, the difference in velocity would not have appeared because of the relatively high velocity found in most glacial sediments.

The first example of the effect of water saturation on seismic velocity was found at station 88, in sec. 9, T. 4 N., R. 9 W., Madison Co., Illinois. The shot hole was augered in alluvial silts and sands forming the bank of the Mississippi River at a point about twenty feet from the water's edge. The bottom of the 6.5-foot hole filled with water before the dynamite was inserted. The velocities in the unconsolidated material were 966 ft./sec. for the unsaturated surficial material and 5179 ft./sec. in the water-saturated material. Velocities in unsaturated material just below the low-velocity aerated layer ranged from 1400 to 1700 ft./sec. at nearby stations which were higher and farther from the river.

The second instance in which the effect of water was noted was at stations 145 and 146 on the flood plain of the Ohio River in the town of Rosiclare in sec. 5, T. 13 S., R. 8 E., Hardin Co., Illinois. A drainage stream for mine water kept the adjacent ground in a swampy condition. Stations 145 and 146 were overlapping profiles, and one shot from each profile was made in wet ground near the stream. Below a surface-layer velocity of 1200 ft./sec., velocities of 5208 ft./sec. and 5435 ft./sec. were recorded from the saturated zones. Eleven other shots made in nearby drier ground showed the average velocity of the dry material immediately below the surface or aerated layer to be 1833 ft./sec. indicating the noticeable effect of the water saturation on the material.

The thin, low-velocity aerated zone was found to extend from the surface to an average depth of 7.3 feet throughout the Arcola-Tuscola area. The average velocity in the layer has been calculated to be 1268 ft./sec., with a minimum velocity of 608 ft./sec. and a maximum of 2543 ft./sec.

[27]Horberg, Leland, Pleistocene deposits below the Wisconsin drift in northeastern Illinois: Illinois Geol. Survey Rept. Inv. 165, p. 28, 1953.

[28]MacClintock, Paul, Recent discoveries of pre-Illinoian drift in southern Illinois: Illinois Geol. Survey Rept. Inv. 19, pp. 26-57, 1929.

CHAPTER V — GEOLOGY OF THE ARCOLA–TUSCOLA AREA

Physiographic Location

The Arcola-Tuscola area lies within the Bloomington ridged plain,[1] which is a part of the Till Plains section of the Central Lowland province.[2] Most of the Wisconsin moraines of Tazewell age are included within the Bloomington ridged plain. Materials deposited by the Shelbyville and Cerro Gordo ice sheets of the Tazewell substage form the surficial deposits within the Arcola-Tuscola area, but the Cerro Gordo moraine is the only moraine in the area. The locations of the Shelbyville and Cerro Gordo moraines are shown in figure 7.[3] No relief features are large enough to require elevation corrections in seismic results. The Kaskaskia River and its tributaries and the Cerro Gordo moraine comprise the only noticeable deviations from the gently rolling plain.

Pleistocene Geology

The Pleistocene deposits in the Arcola-Tuscola area are for the most part of Illinoian and Wisconsin ages. Deposits of Kansan age have been reported in the area by Horberg.[4] No Nebraskan deposits are known, but Foster and Buhle[5] have suggested a possible occurrence in a deep test hole on the University of Illinois campus in Champaign eight miles north of the Arcola-Tuscola area. A stratigraphic column for the Pleistocene of northeastern Illinois by Horberg[6] is given in figure 8.

Leverett[7] mapped and named the Shelbyville and Cerro Gordo moraines and dated them as "Early Wisconsin." Leighton[8] later included them in the Tazewell substage, which is the equivalent of Leverett's Early Wisconsin. Materials deposited by the Shelbyville and Cerro Gordo ice lobes rest successively upon Illinoian drift.

In many places within the area a weathered soil zone separates the Illinoian and Wisconsin drifts for which the name "Sangamon" was proposed by Leverett,[9] who interpreted it as representing an interglacial stage. Horberg[10] has mapped the buried Sangamon plain. The elevation of the surface lies between 600 and 650 feet, as shown in figure 9. The map which was published after the seismic data were collected, corroborates the information gained from the seismic research (see chapter VIII). Foster[11] has published a more detailed study of the Shelbyville-Sangamon contact in the vicinity of Mattoon in Coles County. The northern third of Foster's area is in R. 7 and 8 E., T. 13 N., which is included within the Arcola-Tuscola area.

The buried Sangamon plain in the southern part of the Wisconsin glaciated area (fig. 9) has been described by Horberg[12] as being similar to the flat till plain on the Illinoian drift to the south. The similarity is especially evident in the Mattoon area. Valleys formed in Sangamon time appear to have determined at least in part the present courses of major streams in the area, which includes the Kaskaskia River. Horberg[13] has stated that the preservation of the Sangamon soils in the marginal areas of Wisconsin glaciation testifies to the lack of erosive powers of the Tazewell glaciers. Weathered zones outside the Wisconsin boundary are not noticeably thicker. The lack of erosive power of the ice is attribut-

[1]Leighton, M. M., Ekblaw, G. E., and Horberg, Leland, Physiographic divisions of Illinois: Jour. Geology, vol. 56, no. 1, p. 24, 1948; reprinted as Illinois Geol. Survey Rept. Inv. 129, 1948.
[2]Fenneman, N. M., Physiography of eastern United States: New York. McGraw-Hill, pp. 499-518, 1938.
[3]Leighton, M. M., Ekblaw, G. E., and Horberg, Leland, *op. cit.*, p. 22.
[4]Sample studies, Illinois Geol. Survey files.
[5]Foster, J. W., and Buhle, M. B., An integrated geophysical and geological investigation of aquifers in glacial drift near Champaign-Urbana, Illinois: Econ. Geology, vol. 46, no. 4, p. 379, 1951; reprinted as Illinois Geol. Survey Rept. Inv. 155, 1951.
[6]Horberg, Leland. Pleistocene deposits below the Wisconsin drift in northeastern Illinois: Illinois Geol. Survey Rept. Inv. 165, p. 11. 1953.
[7]Leverett, Frank, The Illinois glacial lobe: U. S. Geol. Survey Mon. 38, pp. 191-221, 1899.
[8]Leighton, M. M., The naming of the subdivisions of the Wisconsin glacial age: Science, vol. 77, p. 168, 1933.
[9]Leverett, Frank, The weathered zone (Sangamon) between the Iowan loess and Illinoian till sheet: Jour. Geology, vol. 6, p. 176, 1898.
[10]Horberg, Leland, *op. cit.*, p. 26.
[11]Foster, J. W., Major aquifers in glacial drift near Mattoon, Illinois: Illinois Acad. Sci. Trans., vol. 44, pp. 85-94, 1951; reprinted as Illinois Geol. Survey Circ. 179, 1952.
[12]Horberg, Leland, *op. cit.*, p. 36.
[13]*Ibid.*, p. 38.

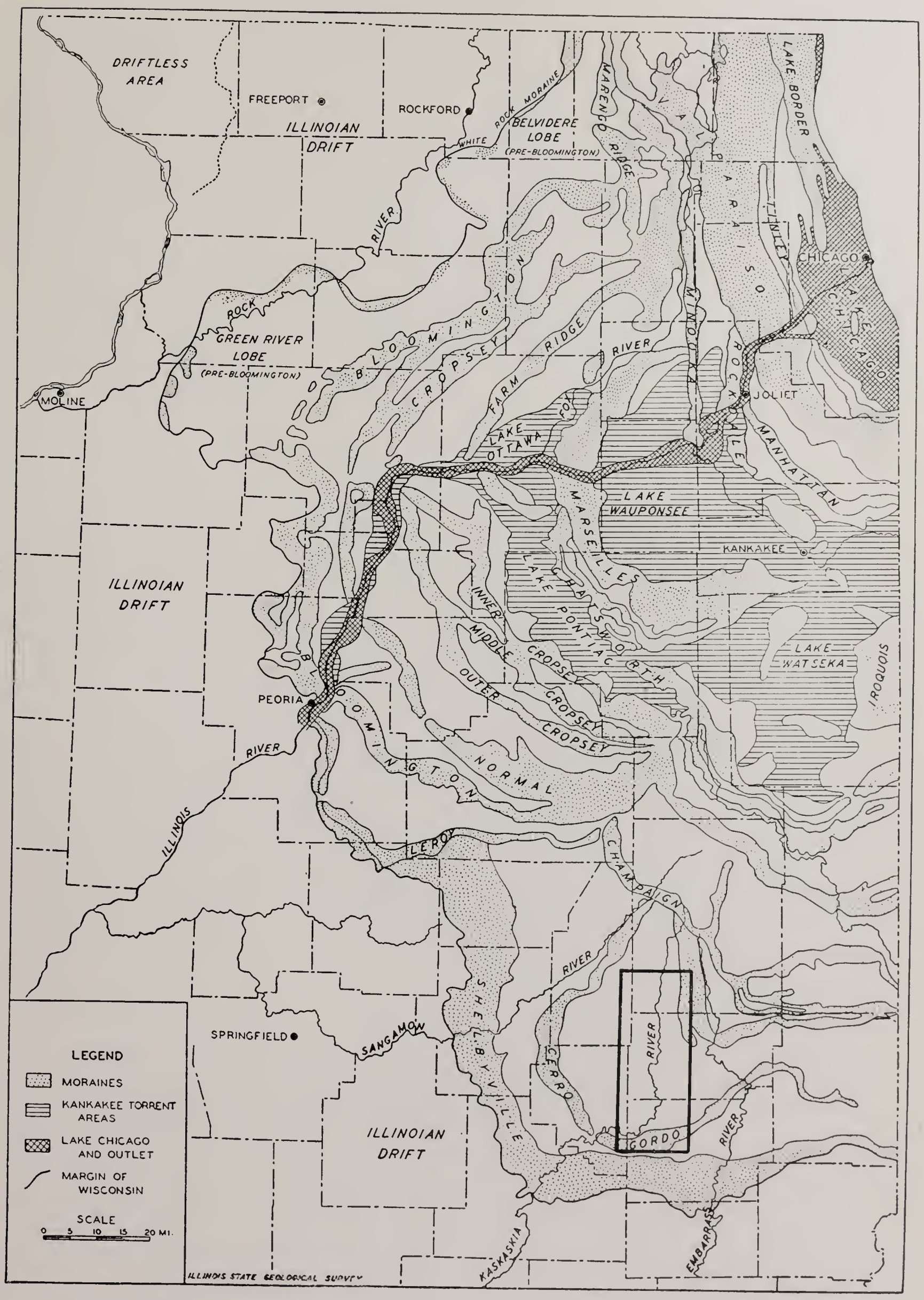


FIG. 7.—Glacial geology in northeastern Illinois. Compiled by George E. Ekblaw from data furnished by the Illinois Geological Survey, January 1, 1942.

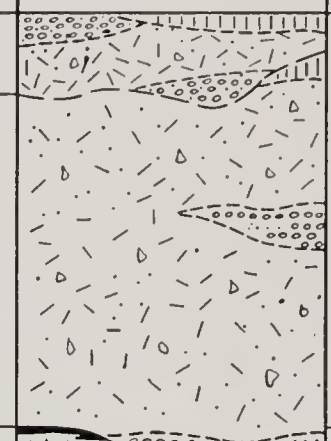
STAGE	SUBSTAGE	SECTION	THICKNESS AVER.	THICKNESS MAX	MATERIAL	ORIGIN
WISCONSIN	CARY		100	300	Till, grovel, sond, silt, loess	Glociol moroines, outwosh deposits, wind-blown loess
	TAZEWELL					
	IOWAN		2	10	Loess	Wind deposit
	FARMDALE		2	12	Silt, loess, peot	Wind, streom, pond, ond swomp deposits
ILLINOIAN	Songomon soil profile		75	150	Till, grovel, sond	Weothered zone Glociol moroines, outwosh
	LOVELAND		5	50	Silt, peot	Streom, pond, swamp, and ? wind deposits
KANSAN	Yormouth soil profile		50	100	Till, grovel, sond	Weothered zone Glociol moroines, outwosh
	PRO-KANSAN ?		5	50	Silt, peot	Streom, pond, swomp, ond ? wind deposits
NEBRASKAN ?	Aftonion soil profile		50	75	Till, grovel, sond	Weothered zone Glociol moroines, outwosh
SANKOTY (MAHOMET) SAND			100	300	Sond, grovel, silt	Streom deposits, probobly outwosh
BEDROCK						

FIG. 8.—Graphic section of Pleistocene deposits in northeastern Illinois. (Horberg, Leland, Pleistocene deposits below the Wisconsin drift in northeastern Illinois: Illinois Geol. Survey Rept. Inv. 165, fig. 1, 1953.)

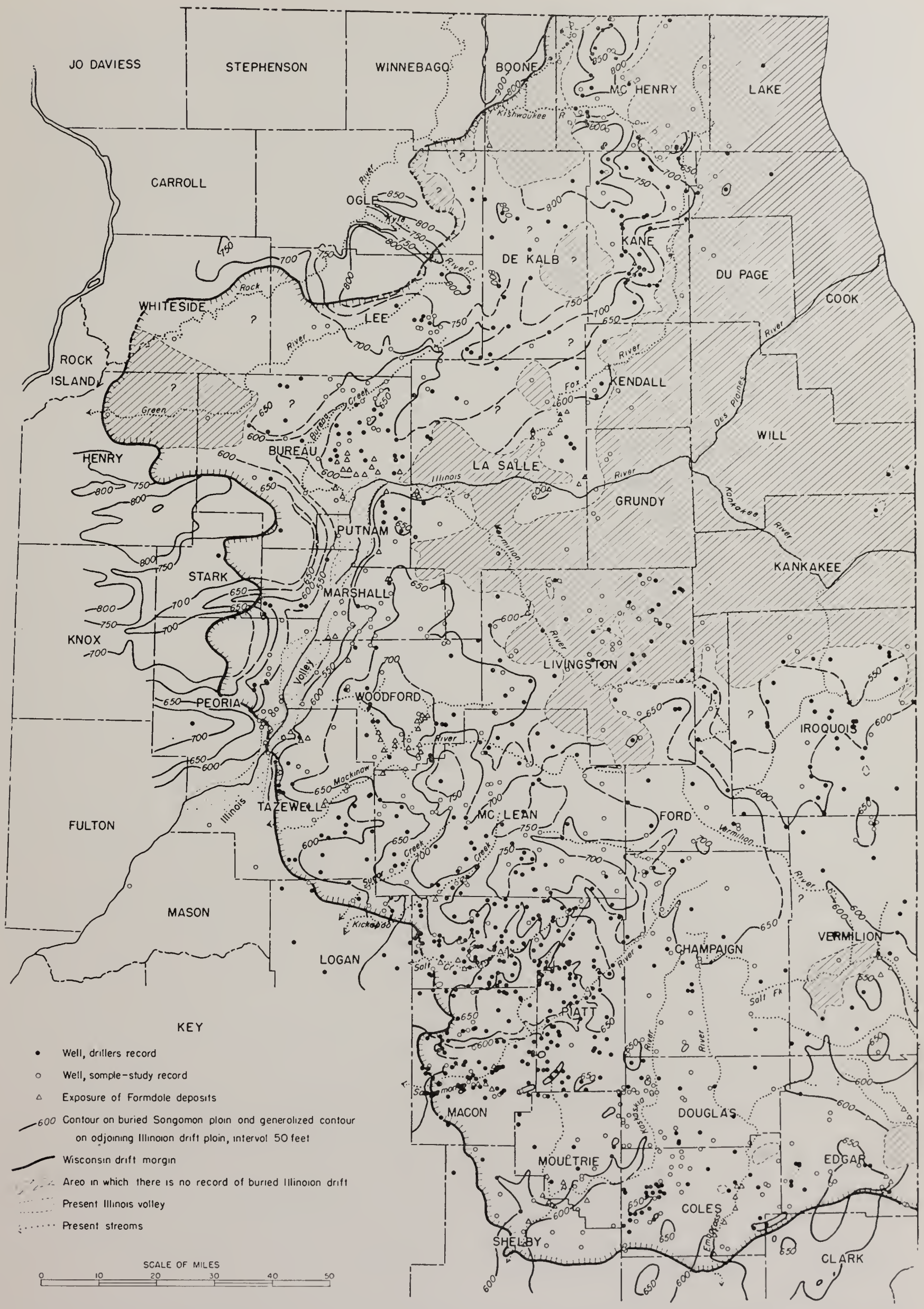


FIG. 9.—Areas underlain by Illinoian and Farmdale deposits and contour map of buried Sangamon plain. (Horberg, Leland, Pleistocene deposits below the Wisconsin drift in northeastern Illinois: Illinois Geol. Survey Rept. Inv. 165, fig. 8, 1953.)

able to the relatively high-level pre-Wisconsin topography, overloading of the ice with debris, and thinning of the ice near the margin.

No occurrences of Farmdale deposits are recorded in well logs from the area, but it is possible that some of the peat zones are of Farmdale age.

All available well data which are applicable to the problem of delineating drift surfaces within the area were used in preparing plates 1 and 2. The data include sample studies which, with few exceptions, were made by Horberg and driller's logs which have been examined and correlated by the writer using recognized criteria for the differentiation of drift sheets. Only those driller's logs were used which indicated a reasonable amount of detail and which were from wells drilled by men whose descriptions were found to be good by sample study comparisons. Exceptions are logs of holes which were bottomed in the well-developed, widespread peat deposit of reported Sangamon age.

Flint[14] has presented several criteria for the differentiation of drift sheets. The most recent listing of criteria, by Horberg[15] is as follows: (1) weathered zones between unweathered drift sheets; (2) extensive buried peat, wood, and other organic matter; (3) deposits of loess and fossiliferous silt between tills; (4) relative positions of buried weathered zones and peat beds with reference to depth, elevation, stratigraphic interval, and distance above bedrock; (5) stratigraphic succession of till, outwash, and other deposits; and (6) differences in physical properties of various drift sheets. Of Horberg's criteria, numbers 1, 2, 4, and 5 have proved to be the most useful in correlating driller's logs in the Arcola-Tuscola area.

Three weathering zones in the upper portion of each of the Nebraskan and Kansan tills in Iowa have been described by various workers. Kay and Apfel,[16] in 1929, divided the upper zone into (1) an oxidized and leached till, (2) oxidized and unleached till, and (3) fresh till. In Illinois, MacClintock[17] used the more deeply leached soil profiles on the pre-Illinoian drift as evidence of a prolonged period of weathering before the deposition of the overlying Illinoian till. The present standard mature profile of weathering for glacial till, proposed in 1930 by Leighton and MacClintock,[18] consists of five horizons numbered from the surface down as follows: (1) surficial soil; (2) chemically decomposed till, constituted chiefly of alteration products and resistant elements of the original till strikingly unlike the original till; (3) leached and oxidized till, otherwise little altered; (4) oxidized unleached till; and (5) unweathered till. This profile is of value in studying driller's logs because soil zones are occasionally referred to and oxidized zones are often reported. Kay and Apfel[19] have commented on the yellow, red, and brown colors resulting from the oxidation of iron compounds in the drift. Drillers often log such zones as "yellow clay," "brown clay," "yellow hardpan," and "yellow drift." Green zones resulting from the reduction of iron compounds are also important and are often reported by drillers as "green clays."[20] Leverett[21,22,23] recognized the importance of soil zones and colored oxidized zones in distinguishing Pleistocene deposits of different ages.

Peat zones encountered in drilling are almost always recorded in reliable driller's logs. They are helpful in tracing stratigraphic units wherever their occurrence is widespread, as in the Mattoon area. In most of the Arcola-Tuscola area, however, there are only scattered references to peat in driller's logs, but where there are references the relative positions of peat beds and weathered zones have been found to be

[14]Flint, R. F., *op. cit.*, pp. 190-194.
[15]Horberg, Leland, *op. cit.*, pp. 8-9.
[16]Kay, G. F., and Apfel, E. T., The pre-Illinoian Pleistocene geology of Iowa: Iowa Geol. Survey, vol. 34, pp. 162-178, 235-252, 1929.
[17]MacClintock, Paul, Recent discoveries of pre-Illinoian drift in southern Illinois: Illinois Geol. Survey Rept. Inv. 19, pp. 26-57, 1929.
[18]Leighton, M. M., and MacClintock, Paul, Weathered zones of the drift sheets of Illinois: Jour. Geology, vol. 38, pp. 31-37, 1930; reprinted as Illinois Geol. Survey Rept. Inv. 20, 1930.
[19]Kay, G. F., and Apfel, E. T., *op. cit.*, pp. 139-141, 218-219.
[20]Horberg, Leland, *op. cit.*, p. 10.
[21]Leverett, Frank, *op. cit.*, pp. 120, 125, 185-187, 1899.
[22]Leverett, Frank, *op. cit.*, pp. 176-180, 1898.
[23]Leverett, Frank, The weathered zone (Yarmouth) between the Illinoian and Kansan till sheets: Jour. Geology, vol. 6, pp. 238-239, 1898.

quite consistent in some areas.[24] In the vicinity of Arcola the positions of soil and weathered zones have aided in the seismic interpretation.

Persistent sand and gravel beds have also assisted in the correlation of driller's logs. Changes in the till colors reported by drillers are significant indicators of differing ages in drift sheets. No one criterion may be used safely to make correlations, but rather a closely integrated use of as many as possible is necessary.

Bedrock Geology

In the Arcola-Tuscola area the rock comprising the bedrock surface is of Pennsylvanian age except where rocks of Mississippian and Devonian age are reported to reach the bedrock surface along the LaSalle anticline.[25,26] In these two quadrangles, the two older systems occur only east of a line extending northward and southeastward from a point about a mile west of Tuscola. Seismic rock velocity determinations indicate that stations 11 and 12 are within this area and that stations 16 and 17 may be included, although the velocities recorded at the latter two stations are more in accordance with Pennsylvanian sandstone or shale velocities recorded elsewhere in the area.

[24]Horberg, Leland, *op. cit.*, p. 10.

[25]Weller, J. M., et al., Geologic map of Illinois: Illinois Geol. Survey, 1945.

[26]Sample studies, Illinois Geol. Survey files.

CHAPTER VI — EQUIPMENT AND FIELD PROCEDURES

Equipment

Modern refraction seismic equipment has been improved to the extent that instruments designed for shallow investigations can be completely portable. As a result, exploration may be carried on in areas which heretofore were inaccessible. The small number of men necessary for field operation and the small original cost of the equipment make seismic surveys by small companies and governmental agencies economically practicable.

The basic equipment used in shallow refraction seismic exploration is shown in figure 10. It is a Century Model 404 12-trace portable refraction seismograph designed and constructed by the Century Geophysical Corporation, of Tulsa, Oklahoma. Earth vibrations caused by a dynamite blast are received by geophones or seismometers which are placed on the surface of the ground. Six of the twelve cylindrical geophones used are shown in figure 10. The earth motion received by the geophones is converted by induction into electrical energy, which is conducted to amplifiers through a 650-foot, 24-conductor cable shown on a back-pack reel at the left in figure 10. The amplifier case is next to the reel. It contains twelve separate manually controlled amplifiers for the twelve geophones. Additional equipment incorporated within the amplifier case includes a.c. balance potentiometers, circuit-continuity test switches, and circuit leakage-to-ground test switches. The amplified electrical impulses travel to the oscillograph, to the right of the amplifier in figure 10. The electrical energy is translated into motion by twelve moving-coil d'Arsonval galvanometers mounted in the oscillograph case. By means of an optical system, the motion of each galvanometer is recorded on

Fig. 10.—Portable shallow refraction seismic equipment.

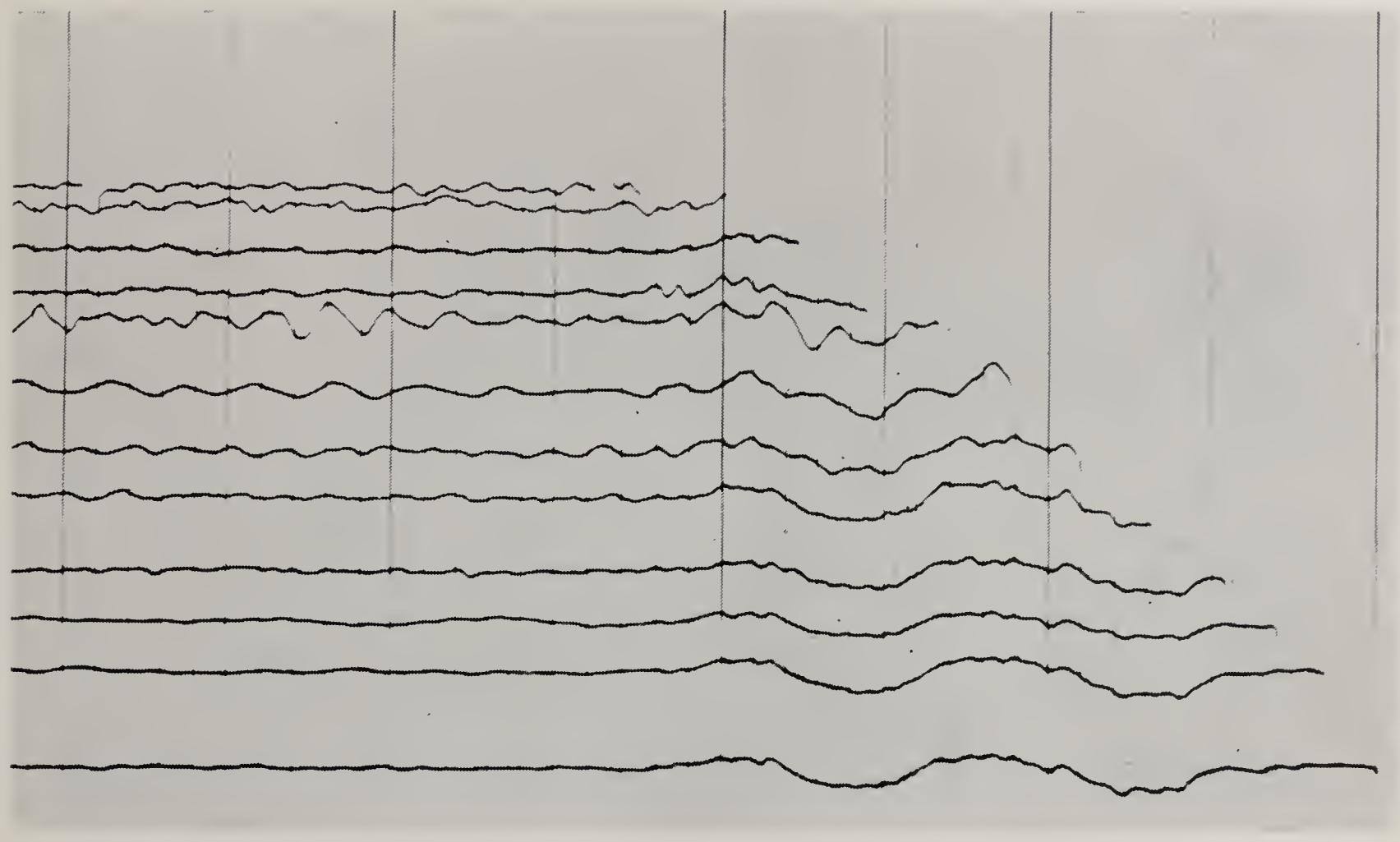


FIG. 11.—Photographic shot record made at station 84A, sec. 5, T. 16 N., R. 8 E., Tuscola quadrangle.

photographic paper which travels at a constant speed through the camera portion of the oscillograph. Timing lines 1/100 second apart are recorded on the moving paper by light which flashes through slots in a disc driven at a constant speed. The exposed photographic record is developed in a daylight developing compartment which forms the lower half of the oscillograph unit. The record of the shot made at station 84A, sec. 5, T. 16 N., R. 8 E., is reproduced as figure 11. The reel next to the oscillograph in figure 10 holds the wire used to fire the dynamite charge. Later modifications have reduced its size considerably.

Several sources of electrical energy are required by the seismograph. A 6-volt storage battery is required to operate the timing motor and light sources in the oscillograph and to heat the vacuum tube filaments in the amplifiers. Dry-cell batteries enclosed in the amplifier case provide the 90-volt plate voltage used by the amplifier tubes. An independent 6-volt source for firing the electric blasting caps which detonate the dynamite is necessary for if the 6-volt storage battery that operates the amplifiers is used, transient a.c. pulses are fed back to the amplifiers when a shot is fired. The original internal 90-volt plate supply proved to be inadequate and a more reliable external 90-volt supply was employed.

Constant handling and moving of the seismograph results in excessive wear on connecting cables and in frequent maladjustment of electrical and mechanical components. Since most of the shot locations in the Arcola-Tuscola area can be reached by automobile the seismograph was shock-mounted to a table built in a station wagon, using easily disconnected fasteners. This greatly reduced maintenance and made possible more efficient field operation. Figure 12 shows the installation and the writer, about to fire a dynamite charge. The modified shot cable reel is mounted on the floor under the table. The large reel on the tailgate contains a 1950-foot cable, which is used in place of the shorter 650-foot cable to record greater depths.

Dynamite and electric blasting caps are the expendable part of seismic equipment. Most of the shots for this investigation were made with semi-gelatin dynamite in which some of the nitroglycerin was replaced by ammonium nitrate. Static-resist-

ant electric blasting caps equipped with 20-foot leg wires were used to detonate the dynamite.

Field Procedures

The methods of operation employed in the field depend on the various applications of the refraction seismograph. When the goal is the determination of depths, operations require a linear arrangement of geophones on the surface of the ground, commonly referred to as a profile. The total distance between the shot point and the farthest geophone in the profile is the geophone spread. The refraction method necessitates a geophone spread of approximately three times the depth being determined.[1]

The degree of accuracy of depth calculations desired is the deciding factor in the spacing of the geophones along the profile. Close spacing is often necessary to prevent interpretive errors. Either of two cables may be used to carry electrical impulses from the geophones to the amplifiers: a 650-foot cable which has geophone attachment contacts spaced at 50-foot intervals and a 1950-foot cable which has contacts at 150-foot intervals. A variety of spacing arrangements may be made with these cables.

Improper placement of the geophones on the ground may seriously affect the quality of seismic records.[2] It was our practice to place the geophones firmly on ground as nearly uniform as possible along the entire geophone spread. Sand, gravel, and various road-surfacing materials were avoided because of their erratic effect on elastic-wave arrival times at the surface. Whenever geophones were located in grassy or plowed areas, holes of equal depth were dug to fresh, compact soil to permit the best reception of ground movement. Geophones were placed below the surface in this manner on windy days, regardless of surface

[1]Linehan, D., Seismology applied to shallow zone research: Am. Soc. for Testing Materials Spec. Tech. Pub. 122, p. 162, 1952.

[2]Washburn, H., and Wiley, H., The effect of placement of a seismometer on its response characteristics: Geophysics, vol. 6, no. 2, pp. 116-131, 1941.

Fig. 12.—Seismograph mounted in station wagon.

materials, to minimize the effect of the wind and wind-blown objects on the geophones. Operations near woods, cornfields, and fences were curtailed in windy weather because of earth vibrations produced by the wind.

The only source of earth-motion energy used in this study was dynamite. Usual procedure is to place one to six sticks of dynamite in a hole made with a two-inch diameter soil auger. Damage to the ground surface has been largely prevented by using shot holes six feet deep for charges of three sticks and less, and holes nine feet deep for more than three sticks. A stick of dynamite containing an electric blasting cap is put into the hole last to prevent the possible expulsion of unexploded dynamite from the hole. Water is the tamping agent used for confining the dynamite charge in the hole. Other tamping materials such as dirt are often used, but they present added hazards in the event of a misfire.

The area in which the field work was conducted is artificially drained by a random network of shallow ceramic drainpipes or tiles, which are easily broken by a dynamite blast. The farmers lay tiles along the road ditches to act as main drains for the tiles which drain the fields. As most of this seismic work has been conducted along the roads, to prevent possible damage to private property, the presence of drain tiles in the ditches has required that many localities be left unexplored.

CHAPTER VII — CALCULATION AND INTERPRETATION OF SEISMIC DATA

Calculation Procedures

For an understanding of calculation procedures, ray-path diagrams are used rather than the wave-front diagrams which were described in chapter II. A ray, because it is drawn perpendicular to a wave front, shows the changes that occur when energy comes from one medium into another in which there is a different seismic velocity. In figure 2,[1] the lines OA, AG, and GH form the ray path for the initial energy arriving at point H on the surface. This is the ray path followed by the initial energy arriving at the geophones on the surface and is the one considered in depth calculations. The ray AG has been refracted at an angle of 90° from the vertical, in accordance with Snell's law (fig. 13), which states that a ray entering a higher-velocity medium will be bent away from the normal. The converse is also true, and gives rise to erroneous refraction data which will be discussed later in the chapter.

Figure 13 shows the refraction of three rays upon entering the higher-velocity, V_2 layer. The three-arrowed ray has an angle of refraction, r, of 90°. If the angle of incidence, i, were greater, no refraction would occur, and all the energy would be reflected back into the V_1 medium. The angle i which results in a refracted angle of 90° is called the "critical angle." This angle is a function of the ratios of the sines of the angles of incidence and refraction and the velocities, as shown in figure 13. When the angle r is 90° its sine function is equal to unity, and at the critical angle sine i = V_1/V_2. The time it takes energy

[1] *After* Leet, L. D., Practical seismology and seismic prospecting: New York, Appleton Century, p. 117, 1938.

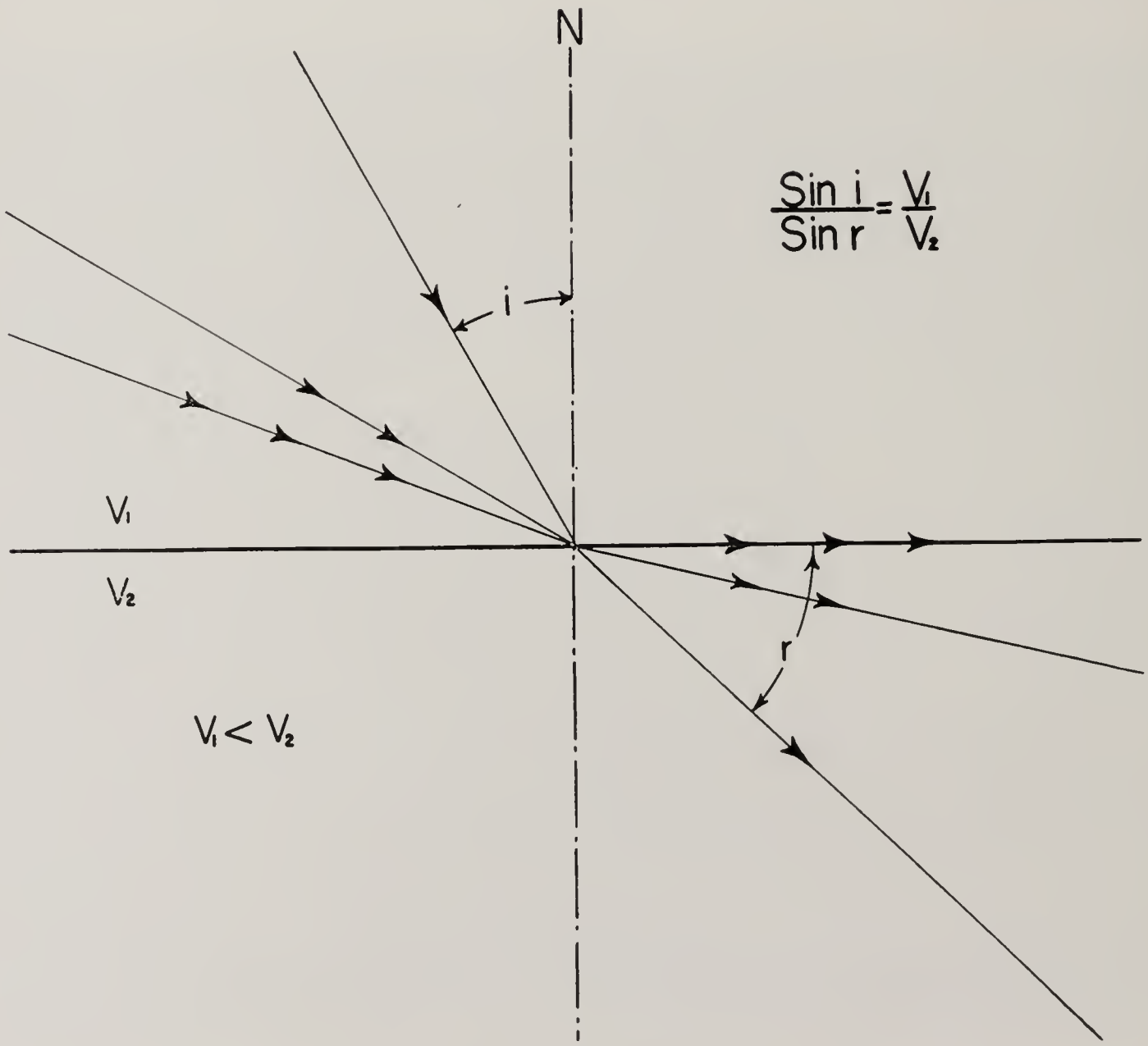


FIG. 13.—Snell's law.

to traverse the path OAGH in figure 2 may be stated as follows:

$$t\ (\text{time}) = \frac{AG}{V_2\ (10{,}000)} + \frac{OA}{V_1\ (5000)} + \frac{GH}{V_1};$$

since

$$OA = GH,\ t = \frac{AG}{V_2} + \frac{2\ OA}{V_1}.$$

The inclined distance OA is equal to the vertical thickness AX divided by the cosine of angle OAX or i in the formula

$$OA = \frac{AX}{\cos i}.$$

The distance AG=OH−2AX tan i. Substituting these quantities in the equation

$$t = \frac{AG}{V_2} + \frac{2\ OA}{V_1},\ t = \frac{OH}{V_2} + \frac{2\ AX}{V_1 \cos i} -$$

$$\frac{2\ AX \tan i}{V_2} = \frac{OH}{V_2} + \frac{2\ AX}{V_1 \cos i} - \frac{2\ AX \sin i}{V_2 \cos i} =$$

$$\frac{OH}{V_2} + \frac{2\ AX}{V_1 \cos i} - \frac{2\ AX \sin^2 i}{V_1 \cos i} = \frac{OH}{V_2} +$$

$$\frac{2\ AX}{V_1 \cos i}\left(1 - \sin^2 i\right) = \frac{OH}{V_2} + \frac{2\ AX \cos i}{V_1}.$$

In order to obtain the quantities required for the final equation above, a travel-time graph is constructed from field data. On a travel-time graph distances are plotted on the abscissa and times are plotted on the ordinate (see fig. 16). The computer gets from the field notes the distance of each detector or geophone from the shot point, or origin on the graph. The times at which the shot energy arrives at the various geophones are read from the photographic record, and the corresponding distances and times are plotted on the graph. Figure 2 indicates the sequence in which energy from the underlying layers reaches the surface. In this report, different velocity layers are numbered by subscripts, as V_1, V_2, etc., from the surface layer downward. In practice, the segments of a travel-time curve are referred to as "velocities." They are, however, actually reciprocals of velocities, since velocity equals distance divided by time.

A subsurface velocity discontinuity is indicated by a change in the slope in the travel-time curve. Two such breaks corresponding to subsurface conditions may be seen in figure 16. The values of V_1, V_2, etc. may be calculated from each curve segment. The cosine function needed for the depth equation is calculated from the equation for Snell's law at the critical angle:

$$\sin i = \frac{V_1}{V_2}.$$

The value for time t is obtained by extending the velocity segments on the curve to the ordinate and then reading the time intercept on the ordinate. This is equal to reading the time at distance zero on the abscissa; therefore, the quantity OH in the equation is equated to zero, and solving for AX the equation becomes

$$AX = T - \frac{V_1}{2 \cos i}.$$

The quantity AX from figure 2 will be referred to as "h" in the calculation results that follow. When more than two layers of materials are encountered in seismic work, the equation for thickness is expanded so that the thickness of each layer may be calculated. The steps followed in such an expansion are given in a paper by Ewing, Crary, and Rutherford.[2]

Figure 14 shows the paths for direct, refracted, and reflected rays for a two-layer problem. Differentiating the three types of arrivals often presents a problem. As seen in figure 2, direct energy—that which is radiating outward from the point O in the 5000 ft./sec. (V_1) material—will be recorded first until the straight head waves from the lower layers arrive, after which time the faster head waves precede the direct waves and are recorded first. In comparing reflected and refracted energy arriving at any one geophone, reflected energy following the ray path (see fig. 15) travels the entire distance through the low-velocity material V_1. The refracted energy travels through less of the low-velocity material and goes through some distance of

[2]Ewing, M., Crary, A. P., and Rutherford, H. M., Geophysical investigations in the emerged and submerged Atlantic Coastal Plain; pt. 1. Methods and results: Geol. Soc. Am. Bull., vol. 48, pp. 763-764, 1937.

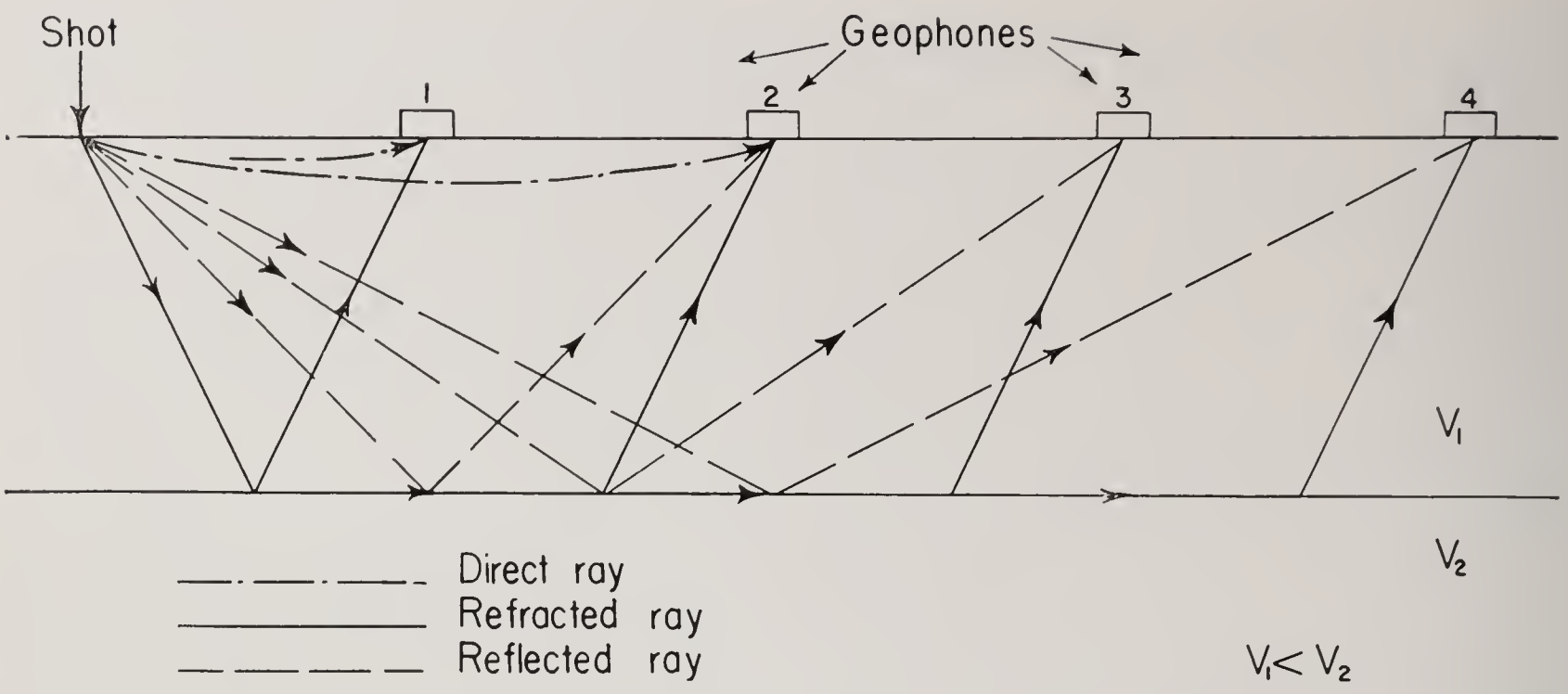


FIG. 14.—Direct, refracted, and reflected ray-path relationship.

the high-velocity material V_2 and as a result arrives at the surface before the reflected energy. The travel times are compared in figure 15.

Sample depth calculations for station 81, NE cor. sec. 15, T. 14 N., R. 8 E., Arcola quadrangle (pl. 1), are given on this page. Velocity and travel time data were taken from the graph shown as figure 16.

The 9804 ft./sec. layer is considered to be bedrock, and to obtain the total depth to bedrock, h_1 and h_2 are added, plus a correction factor consisting of one-half of the depth of the shot hole. Therefore, the total depth to bedrock at station 81 becomes $h_1 + h_2 +$ correction $= 83.5$ feet. A test hole for the town of Arcola was drilled in the SW cor. sec. 11, T. 14 N., R. 8 E. The driller's log of the hole shows that shale was encountered at 83.5 feet.

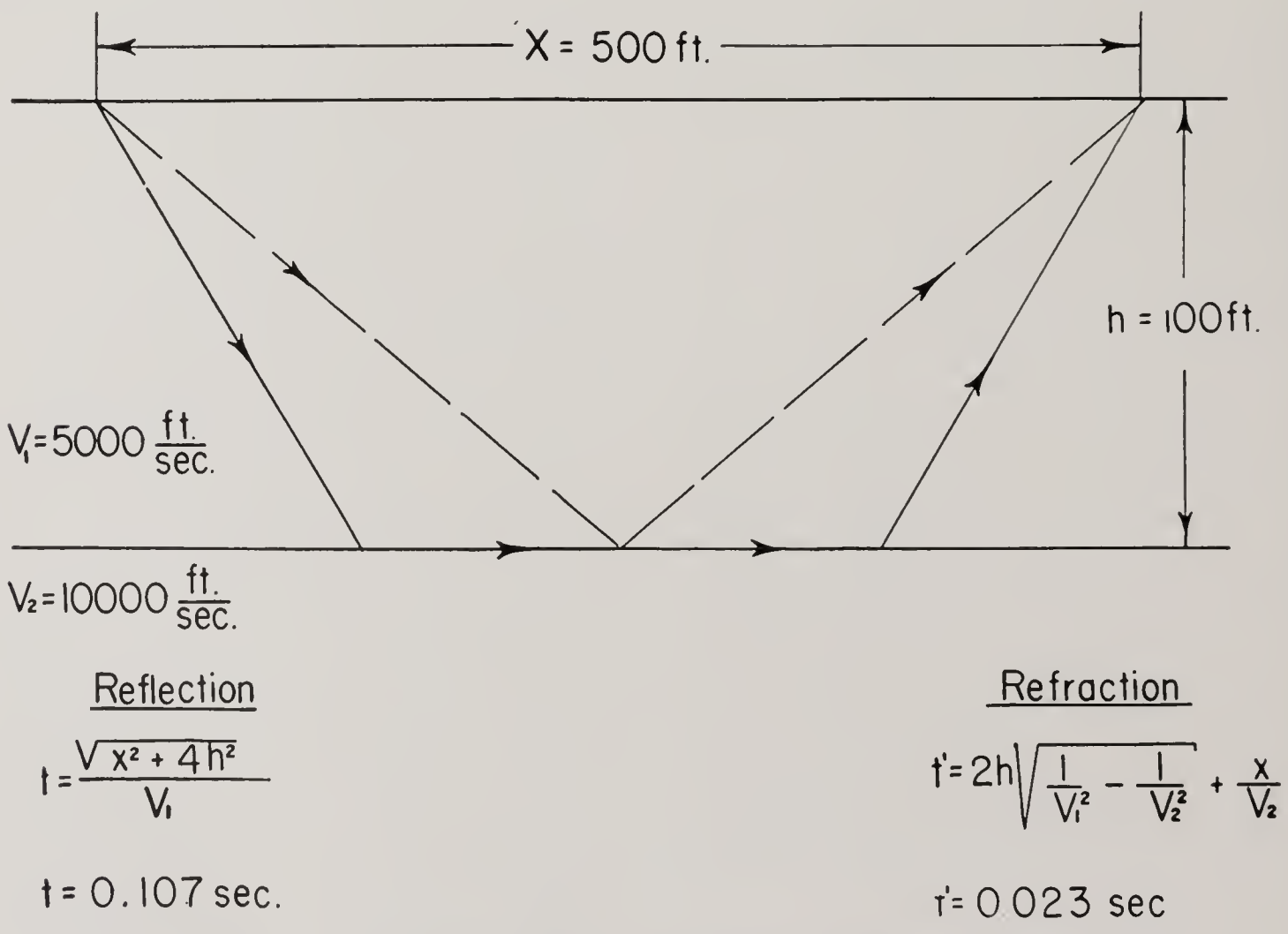


FIG. 15.—Comparison of reflected and refracted ray travel times.

$V_1 = 1371$ ft./sec. $T_2 = .0052$ sec.
$V_2 = 6079$ ft./sec. $T_3 = .0250$ sec.
$V_3 = 9804$ ft./sec.

$$\sin \delta = \frac{V_1}{V_2} = \frac{1371}{6079} = .266;\ \delta = 13.1°;\ \cos \delta = .974$$

$$h_1 = \frac{T_2 V_1}{2 \cos \delta} = \frac{.0052\ (1371)}{2(.974)} = 3.66 \text{ feet}$$

$$\sin \beta_1 = \frac{V_1}{V_3} = \frac{1371}{9804} = .140;\ \beta_1 = 8.1°;\ \cos \beta_1 = .990$$

$$\sin \beta_2 = \frac{V_2}{V_3} = \frac{6079}{9804} = .620;\ \beta_2 = 38.3°;\ \cos \beta_2 = .785$$

$$h_2 = \left[T_3 - \frac{2\ h_1 \cos \beta_1}{V_1} \right] \frac{V_2}{2 \cos \beta_2} = \left[.0250 - \frac{7.32\ (.990)}{1371} \right] \frac{6079}{2(.785)} =$$

$$\left[.0250 - .0053 \right] 3872 = (.0197)\ 3872 = 76.3 \text{ feet.}$$

The symbols used for angles in the above computation are those used by Woollard.[3] The Greek letter used for an angle indicates the interface at which the critical angle occurs and the numerical subscript denotes the interface to which the angle refers (fig. 17). Thus, γ_2 refers to the angle at the second velocity interface, in a four-layer example, on the ray path that produces a critical angle at the third interface.

We have been discussing depth calculations only in horizontally layered beds; in practice, such ideal conditions are seldom encountered, especially in mapping an uneven bedrock surface. When a travel-time graph is constructed from seismic data gathered over underlying dipping velocity interfaces, the velocity segments of the curve do not represent the true velocities of the materials but rather apparent velocities whose quantities are functions of the angles of dip.

The intercept method which was used for the calculation of depths at station 81 has been modified for use with dipping beds. The thicknesses measured by the modified method are those which are perpendicular to the interfaces involved in each calculation, as shown in figure 17. If the dip exceeds some maximum angle, which has been chosen as 8° for this work, the depth determinations will be less than the actual depth. The true thickness of any layer may be obtained by dividing the calculated thickness by the cosine function of the angle of dip, w.

[3]Woollard, G. P., Refraction seismic lecture material: Madison, Univ of Wisconsin, 1950.

Interpretation of Seismic Data

The development of the calculation methods described in the preceding section and the subsequent calculations made with the methods are dependent upon several basic assumptions. The velocities in successive strata should increase as the depth increases; the materials composing the strata should be such that the velocities in any direction are the same for any one stratum; the strata should have sufficient thickness; and the boundaries between the strata should be plane surfaces. Strata having velocities lower than velocities of the overlying strata or strata which are too thin do not appear in a travel-time curve.[4] When any of these assumptions is not fulfilled, depth determinations contain errors which may be of considerable magnitude.

When seismic depth determinations differ greatly from those known from nearby control points, as occasionally happens, the cause is believed to be an insufficient in-

[4]Jakosky, J. J., Exploration geophysics: Los Angeles, Trija Pub. Co., pp. 765-766, 1950.

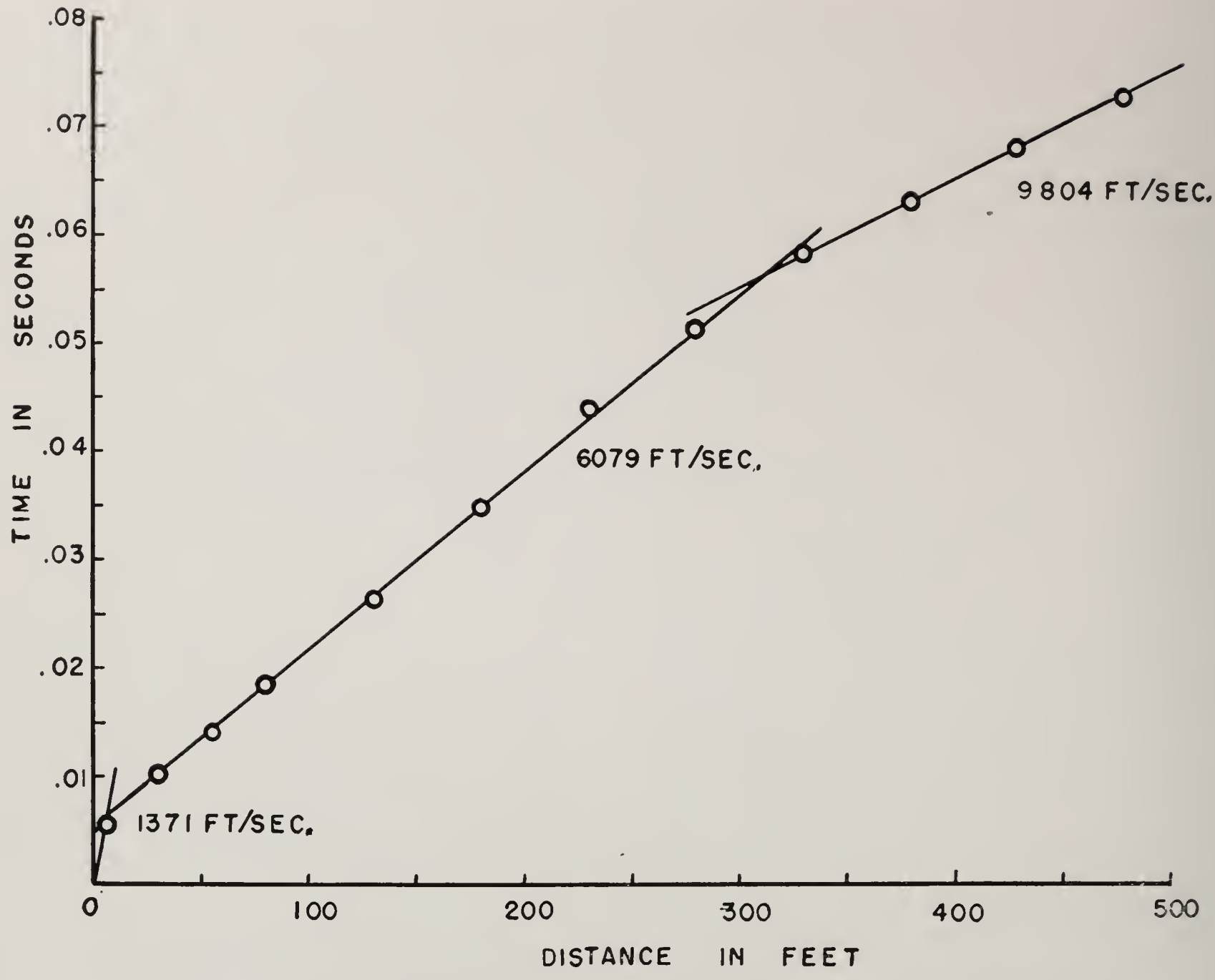


FIG. 16.—Travel-time graph for station 81, sec. 15, T. 14 N., R. 8 E., Arcola quadrangle.

crease in velocity in the bedrock in relation to the overlying unconsolidated material or even a decrease in velocity in some rare cases. The velocities of longitudinal waves overlap considerably when representative velocities from various ages and types of sediments are examined. Recorded velocities higher than 7000 ft./sec. in unconsolidated glacial deposits are higher in some instances than many recorded velocities in shales and sandstones.[5,6] High velocities have been recorded by other writers. Linehan[7,8,9] has referred to compact, older tills having velocities up to 9000 ft./sec. and has called attention to the fact that it is often difficult to determine from velocities alone whether a material is bedrock or a compact till. Shepard and Haines[10] have reported that velocities in excess of 5000 ft./sec. in unconsolidated material usually indicate the presence of compact glacial till. The seismic velocities recorded during the work on the Arcola-Tuscola area are on file at the Illinois State Geological Survey.

Most supposed erroneous bedrock depths in the Arcola-Tuscola area are in places where there is little or no control. An exception to this is in the SW cor. sec. 4, T. 14 N., R. 8 E., where the A end of station 138 was at a test hole (fig. 22). The 7692 ft./sec. V_3 velocity recorded at station 138A could very easily represent sandstone or shale as it is quite high for unconsolidated

[5]Heiland, C. A., *op. cit.*, pp. 468-472.
[6]Birch, F., et al., Handbook of physical constants: Geol. Soc. Am. Spec. Paper 36, pp. 95-98, 1942.
[7]Linehan, D., and Keith, S., Seismic reconnaissance for groundwater development: Jour. New England Water Works Assoc., vol. 63, no 1, p. 82, 1948.
[8]Linehan, D., Seismology as a geologic technique: Highway Research Board Bull. 13, p. 80, 1948.
[9]Linehan, D., Seismology applied to shallow zone research: Am. Soc. for Testing Materials Spec. Tech. Pub. 122, p. 165, 1952.
[10]Shepard, E. R., and Haines, R. M., Seismic subsurface exploration on the St. Lawrence River project: Am. Soc. Civil Eng. Proc., vol. 68, no. 10, p. 175, 1942.

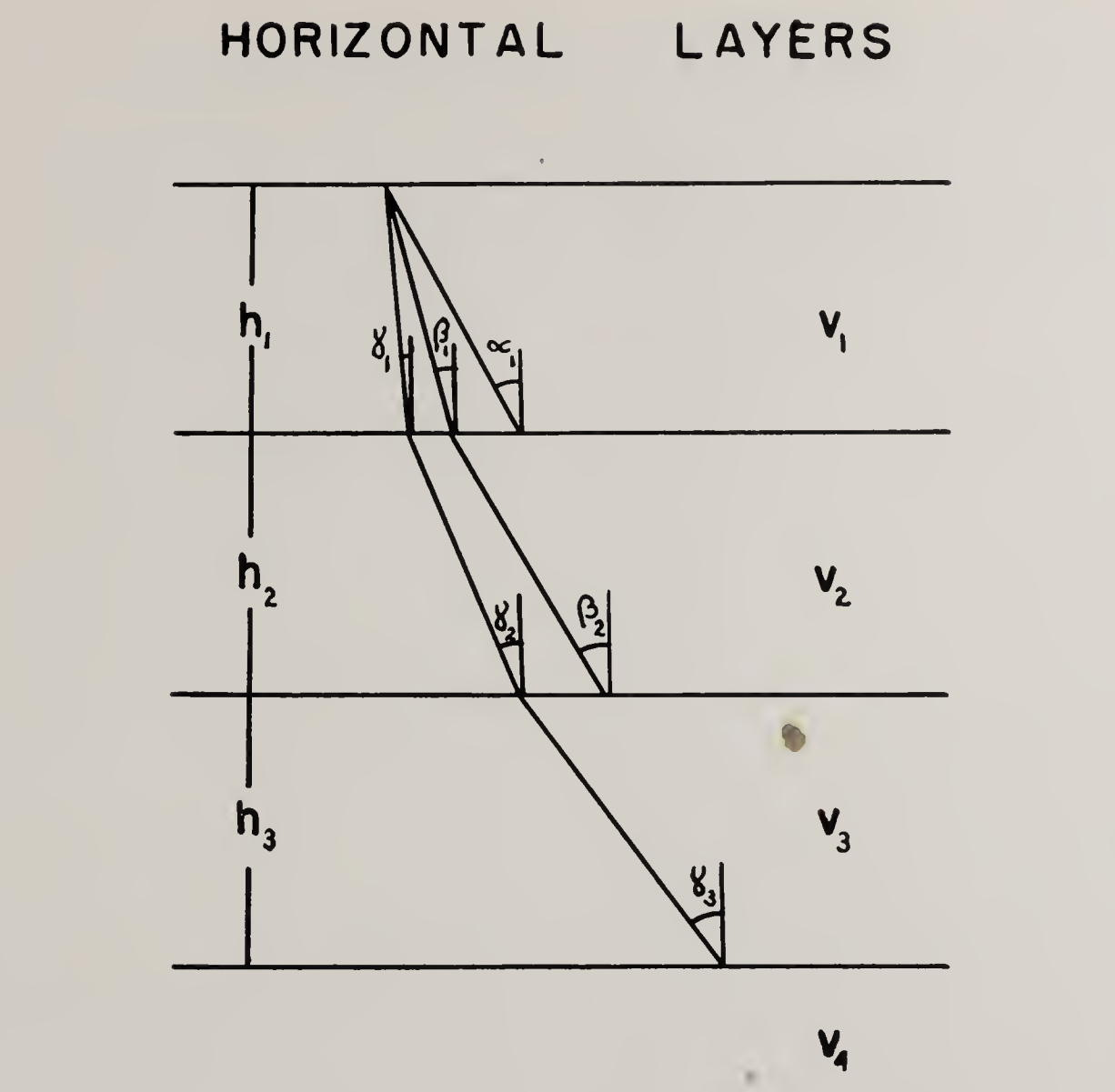


FIG. 17.—Angle notation for horizontal layers and dipping layers.

material in this area. The driller's log data for the test hole is as follows:

E. C. Baker and Sons, Arcola City Test Hole 7-53, SW cor. sec. 4, T. 14 N., R. 8 E., Douglas Co., elev. 671 feet (estimated from topographic map).

	Depth in feet
Soil	0–4
Yellow clay	4–14
Blue clay	14–47
Hardpan	47–58
Blue sand-clay	58–80
Pack gravel	80–84
Mud sand	84–90
Shale	90–93

The calculated depth to the V_2/V_3 interface is 58 feet, which corresponds with the top of the blue sand-clay (till) of the driller's log rather than the bedrock top. No velocity of greater magnitude was recorded at the A end, which indicates that the velocity of the shale was either less or not appreciably greater than that of the till velocity recorded. At the B end of the station the higher-velocity bed indicated is probably a limestone. As there was a time delay on this portion of the graph, the B end is atypical.

Station 122 in the SE cor. sec. 5, T. 14 N., R. 8 E., was shot before a test hole was put in. The depth to the 12,500 ft./sec. layer at the A end was calculated as 151 feet, which was considered to be the top of bedrock. The test well later drilled near the A end indicated that the seismograph had not recorded rock until the refracted energy returned from a deeper layer, here probably limestone. It can be assumed that the conditions are the same as those at station 138 and that the bedrock surface was not recorded because the high till velocity masked the low bedrock velocity. Several erroneous bedrock depths on table 2 are attributable to similar conditions, especially those depths shown at stations 13, 23, and 44. Without a field program for making continuous profiles, there is no way of ascertaining the accuracy of bedrock depth determinations where there are thick glacial sediments in which velocities in the older tills are high and the bedrock is shale. Such a program is not possible in the Arcola-Tuscola area because of the network of buried farm tiles. Selecting the velocity which represents bedrock at any one location often depends upon nearby control and a knowledge of the representative velocities of an area.

It is possible in refraction work even in the ideal situation where velocity increases in succeedingly deeper layers to obtain data which do not record energy from one or more of the layers. Schmidt[11] provided diagrams that indicate that subsurface beds must have a certain minimum thickness in order for a given ratio of velocity to be represented on a travel-time curve as a first arrival of energy. This phenomenon is brought about by the extent to which the velocity in the layer underlying the masked layer exceeds that in the masked layer, thereby offering a correspondingly more rapid path of travel. Wave-front diagrams show how the faster velocity waves reach a point at the surface at the same time as those of the masked layer with the critical thickness. The greater the difference between the two velocities, the greater the critical thickness. If a layer is not indicated on a travel-time curve, the resulting depth computations will be erroneous for all interfaces below the masked layer. The apparent depth calculated from such a travel-time graph for first arrivals is always less than the true depth.

The magnitudes of error for various velocity ratios have been calculated and tabulated by Brinckmeier.[12] The percentage errors are of academic rather than practical value, for the thickness of the masked zone must be known from accurate control data in order to calculate the error correctly. The fact that the examples of masked phases from the Arcola-Tuscola area do not exhibit considerable error is attributable to the relatively low velocity ratio between the layers. Of more practical value is a nomograph constructed by Maillet and Bazerque[13] for the determination of critical

[11]Schmidt, O. von., *op. cit.*, 1931.
[12]Brinkmeier, G., Sonderfälle aus der Praxis des seismischen Verfahrens: Beitr. angew. Geophysik. vol. 4, pp. 154-158, 1934.
[13]Maillet, R., and Bazerque, J., La prospection seismique du sous-sol: Annales des Mines, vol. 20, Douzieme serie, p. 314, 1931.

thicknesses, which may also be found in the writings of Leet.[14,15] The information required is the thickness of the V_1 layer and the angles β_2 and γ_3 (i_1 and i_2 in nomograph), assuming a three-layer problem with the middle V_2 as the masked layer. The velocity in the V_1 layer will not appear on the travel-time curve but may be obtained with relatively good accuracy from nearby seismic records in which it does appear.

The velocities recorded in the Arcola quadrangle appear, from a study of both the seismic data and the well data, to fall into five categories. The first group consists of the aerated zone, with an average velocity of 1397 ft./sec. The second group has an average velocity of 4761 ft./sec., and has been tentatively assigned to the Cerro Gordo age. The third category has an average velocity of 5852 ft./sec.; strata in this category have been tentatively assigned to Shelbyville time. An average velocity of 7187 ft./sec. has been assigned to what are believed to be deposits of Illinoian (undifferentiated) age. The fifth group consists of what are believed to be bedrock velocities, and no average of these has been made.

Within the Arcola quadrangle, some seismic records do not indicate the velocity change that has been assumed to occur at the Wisconsin-Illinoian contact. These records are from an area where the bedrock is at a relatively shallow depth (i.e., 80 feet±) and several are at or near test holes or wells which are reliably logged. All the seismic stations near control points which do not show the velocity change expected at or near the Wisconsin-Illinoian contact have been examined for masked zones.

A representative example is station 79, in the NW cor. sec. 11, T. 14 N., R. 8 E., beside a test hole. The seismic data for this station show a V_1 layer of 1677 ft./sec. velocity and a thickness of 6.0 feet; a V_2

[14]Leet, L. D., *ibid.*, p. 148.
[15]Leet, L. D., Earth waves: Cambridge, Harvard Univ. Press, p. 76, 1950.

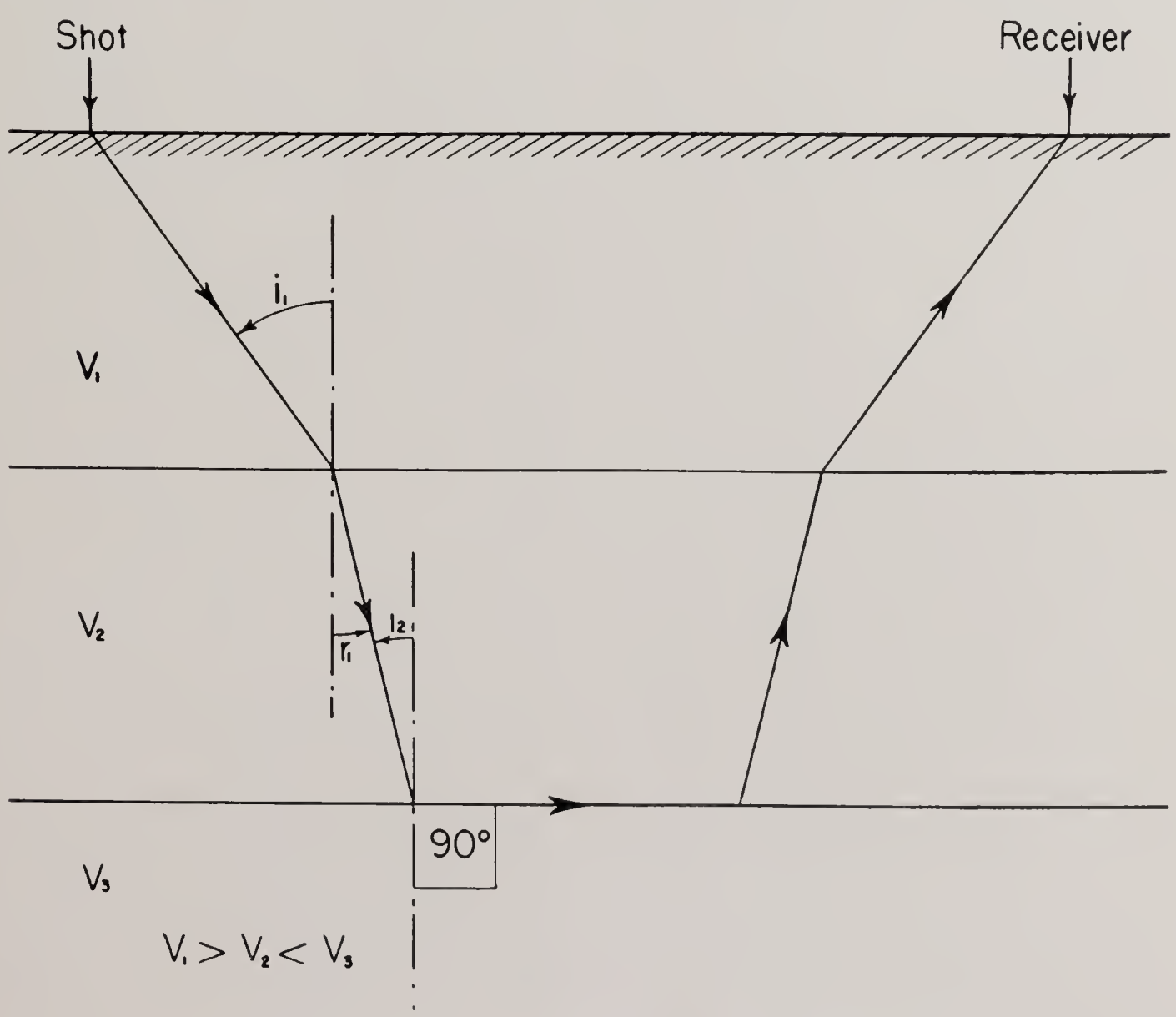


FIG. 18.—Refracted ray paths when velocities do not increase with depth.

layer of 5814 ft./sec. velocity and a thickness of 69 feet; and a bedrock velocity of 11,363 ft./sec. The calculated depth to bedrock was 75 feet and there was no evidence of either the 4761 ft./sec. average velocity of Cerro Gordo deposits or the 7187 ft./sec. average velocity of Illinoian deposits. The well log is as follows:

E. C. Baker and Sons, Arcola City Test Hole 3-53, NW cor. sec. 11, T. 14 N., R. 8 E., Douglas Co., elev. 660 feet (estimated from topographic map).

	Depth in feet
Soil	0–1
Yellow clay	1–12
Blue clay	12–28
Blue sandy clay	28–48
Peat	48–49
Blue sandy clay	49–58
Hardpan	58–72
Yellow hardpan	72–76
Red shale	76–78

The Wisconsin-Illinoian contact is believed to be at 49 feet in the well. With the information from the seismic shot and the well, the following calculations[16] were made:

$V_2 = 5814$ ft./sec.

$V_{2a} = 7187$ ft./sec. (average)

$V_3 = 11{,}363$ ft./sec.

$$\sin \beta_2 = \frac{5814}{7187} = .809 = 54.0°$$

$$\sin \gamma_3 = \frac{7187}{11363} = .633 = 39.3°$$

A thickness of 43 feet for the V_2 layer was arrived at by subtracting the aerated zone thickness of 6 feet from the 49 feet from the log. When the values for β_2 (i_1) and γ_3 (i_2) are inserted in the nomograph, a value of 1.0 is obtained for Y. Y is the number which is equated to the ratio h_{2a}/h_2 to obtain the critical thickness. Therefore, the critical thickness for $h_{2a} = h_1$ (Y) is 43 feet and at least that thickness of material of 7187 ft./sec. velocity would be required to record first arrivals on the travel-time curve. Inspection of the well log shows that there is only 27 feet between the peat layer and bedrock. In the depth to bedrock determination, there was a one-foot difference between the seismic determination and the test hole data. Specific errors cannot be calculated because the depth measured by the seismograph is displaced laterally from the shot point by wave refraction, and therefore the depth measured is never exactly under the shot point.

In sec. 33 of T. 15 N., R. 8 E., along the northern limits of the town of Arcola, stations 141 and 155 were shot in such a way that 141B and 155A are the same point. There was no evidence of a velocity characteristic of Illinoian till in either 141A or B but the travel-time curve for 141B revealed a rather sudden increase in depth to bedrock. For this reason, station 155 was made extending west from 141B. The A end of station 155 confirmed the increase in bedrock depth while the B end showed a return to the bedrock depth found at 141. A velocity characteristic of the Illinoian deposits of the area was found at 155A but not at the shallower B end. From data obtained from the seismic records and the nearby well (see fig. 22), minimum thickness calculations were made. A minimum thickness of 40 feet was found to be required at 155A to be recorded, using the seismic velocities without the aid of an average velocity. A total of 66 feet was present here between the Wisconsin-Illinoian contact and bedrock, and the contact was represented as expected on the travel-time curve for 155A. Since the velocities at 141A and B and 155B were quite close to those at 155A, the thickness of 40 feet was assumed to be needed at these points. Only 30 feet was present at 141A and B and 26 feet at 155B. The depth to the top of the Illinoian in the well is 67 feet, and the depth was calculated from seismic data to be 67 feet at 155A.

Sediments of Illinoian age may not be present in some of the areas of shallow bedrock depths. If they are present but not recorded, they are believed to have been missed seismically either for the reason just discussed or because of the presence of relatively low-velocity sands and gravels. Well-developed deposits of sand and gravel

[16]Maillet, R., and Bazerque, J., *Ibid.*, pp. 313-315.

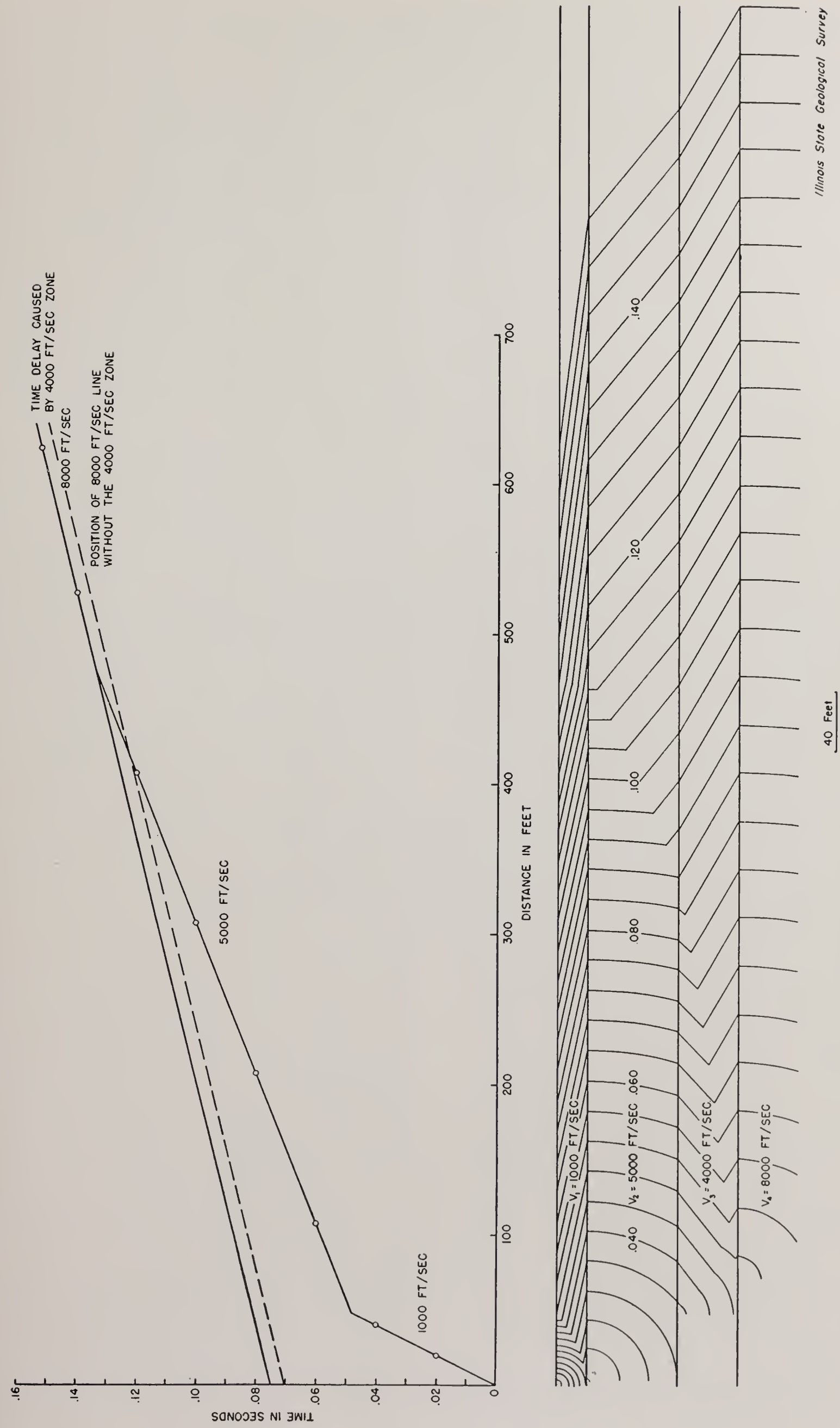


FIG. 19.—Effect of a buried low-velocity layer on a travel-time graph.

give a characteristic delay in the travel-time curve. An examination of seismic records and well logs discloses that the "minimum thickness" concept applies to stations 60, 69, 75, 78, 79, 80B, 81, 83, 110, 125, 140, and 155B in the Arcola quadrangle. The presence of thick sections of low-velocity material or a combination of this and insufficient thickness are assumed to cause the absence on the seismic records of the Wisconsin-Illinoian contact at stations 73, 74, 76, 80A, 139, and 154. The greater number of stations not recording the Wisconsin-Illinoian contact in the Arcola quadrangle compared with the Tuscola quadrangle (plate 2) is believed to be explained by the shallower bedrock surface in much of the Arcola quadrangle and the correspondingly thinner section of Illinoian drift.

The incidence of one layer underlying another layer which has higher velocity can introduce considerable error into depth calculations made from refraction data. Energy from the bedrock surface reaching the interface between two such layers must be refracted toward the normal, according to Snell's law, rather than away from it as required by refraction seismic theory. The ray path of energy that attains a critical angle at the V_2/V_3 interface is shown in figure 18. For such a ray path the only layers to be represented on a travel-time curve are the V_1 and V_3 layers. The resulting depth calculation for the top of the V_3 layer, or what would appear to be the V_1/V_2 interface from the graph, would be greater than the true depth because the total travel time has been increased by some increment dependent upon the thickness of the low-velocity bed and its velocity. A wave-front diagram (fig. 19) has been constructed for a hypothetical case involving a low-velocity bed. The accompanying travel-time graph gives no indication of the presence of the low velocity V_3 layer. If the depth to the V_4 layer (the V_3 layer of the graph) is calculated in the usual manner using the travel-time curve data, the depth is 140 feet, which is 20 feet more than the actual depth.

The wave-front diagram in figure 19 illustrates the theoretical consideration that there can be no indication at the surface of a low-velocity layer. Under certain conditions, however, a low-velocity layer is indicated on the travel-time curve by a discontinuity in some portion of the curve which in effect reveals the increase in the travel time required for waves traversing lower layers. The discontinuity with its lost increment of time may be called a "delay." The travel-time graph for station 139 (fig. 20), in sec. 9, T. 14 N., R. 8 E., exhibits a delay of a sort that usually indicates a fault or a dike in consolidated sediments, and is caused by entirely different conditions.

There are few references in the literature to the phenomenon of delays in unconsolidated materials. Shepard and Haines[17] cite an example of a delay as the result of shooting over frozen ground. Bird[18] also discussed seismic results from shooting over frozen ground, but stated that the graph would have a typical two-layer curve instead of a three-layer curve, showing the low-velocity layer under the temporarily high-velocity frozen layer. The resulting error in depth computation was discussed, but there was no mention of a delay. A delay appeared in a travel-time curve published by Wesley.[19] Drilling of the seismic site gave evidence of an eight-foot layer of low-velocity sand within the section of glacial till. Wesley's depth to bedrock compared very well with the drill data, although no mention was made of the technique used in calculating the depth. A recent paper by Brown and Robertshaw[20] contains a number of examples of delays encountered in their work in unconsolidated sediments. Delays have many times been thought by investigators to be caused by instrument failures or to buried channels.

A problem to be considered is why one part of a segment of the travel-time curve

[17]Shepard, E. R., and Haines, R. M., *op. cit.*, p. 1752.
[18]Bird, P., Experience with geophysics in New York State: Am. Soc. for Testing Materials Spec. Tech. Pub. 122, p. 153, 1952.
[19]Wesley, R. H., Geophysical exploration in Michigan: Econ. Geology, vol. 47, no. 1, p. 61, 1952.
[20]Brown, P. D., and Robertshaw, J., A seismic survey-determination of the thickness of unconsolidated deposits overlying shallow mine workings: Colliery Guardian, vol. 187, no. 4829, pp. 347-353, 1953.

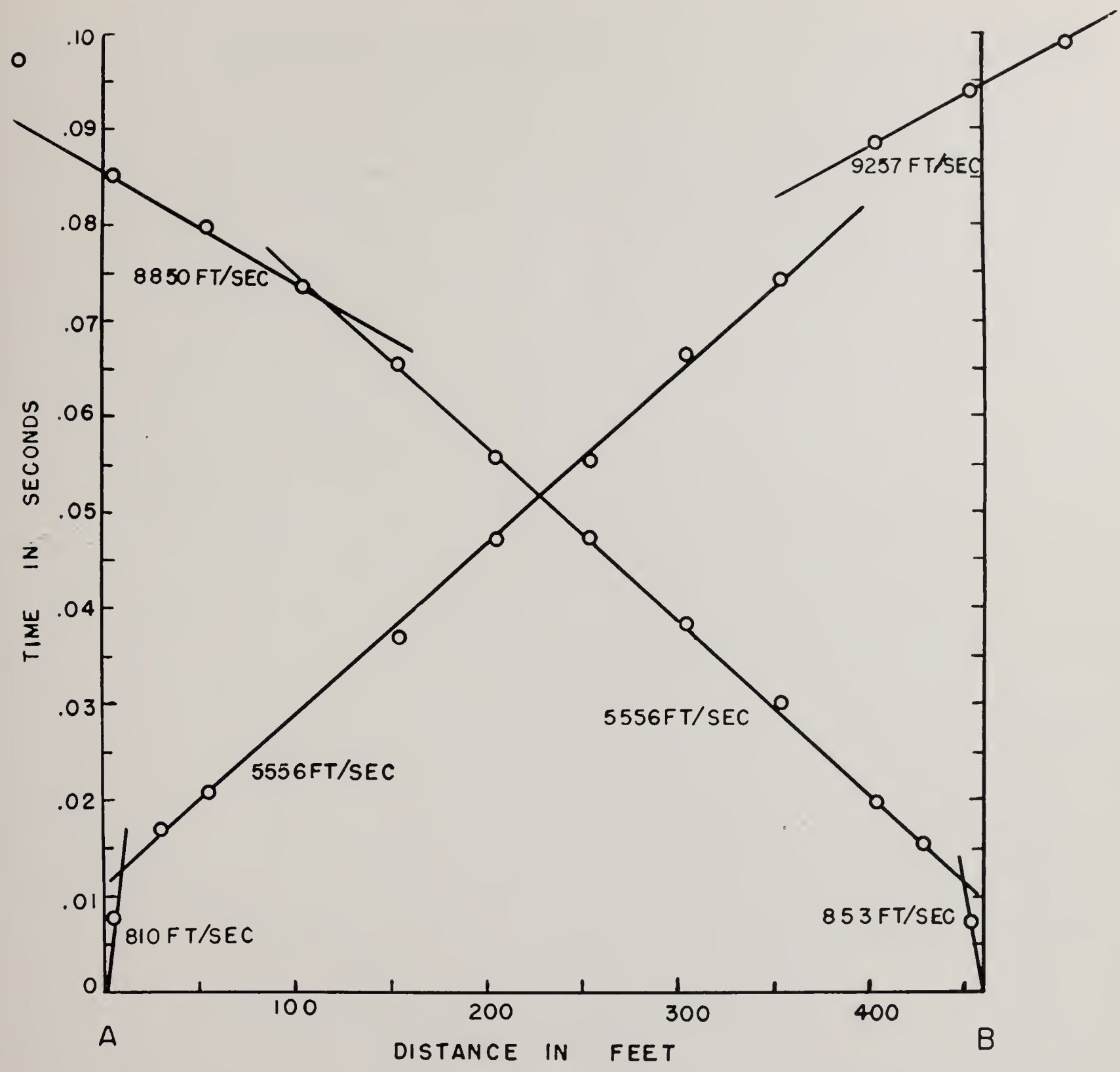


FIG. 20.—Travel-time graph for station 139, sec. 9, T. 14 N., R. 8 E., Arcola quadrangle.

disappears permitting the next faster segment to appear on the graph at a closer than normal distance to the shot point. It is the combination of these two factors that causes the delay. If the velocity segment from the layer immediately overlying the low-velocity bed were to continue unaffected by other conditions, the curve would be the "normal" curve seen in fig. 19. In such a case, the time delay is still present but not seen as a curve discontinuity.

The wave energy refracted into the underlying high-speed bed is not attenuated as the distance from the shot point increases because it is this energy which arrives at the surface as the 8000 ft./sec. layer. The head waves refracting into the low-velocity bed also cannot be attenuated, since they transmit the energy from the underlying bed to the surface, unaffected in strength. Actual shot records exhibit no change in energy magnitude after a delay, which indicates that there has been no attenuation of the energy returning from below the low-velocity bed. Delay is caused by an interference in the propagation of the waves in the overlying V_2 layer so that the head waves coming from the lower high-speed bed are recorded on the surface. Ordinarily, even with a low-velocity bed present, the high-velocity head waves arrive at the surface later than those of the V_2 bed in the wave-front diagram until the coincident time line reaches the surface.

The problem has been discussed with engineers experienced in wave propagation theory and research, who believe that the discontinuity in the travel-time curve is caused by a wave interference phenomenon resulting from a certain combination of wave frequencies, wave lengths, and thicknesses of the layers.[21,22] The many unknown factors, such as the true nature of the interfaces between layers and the wave frequency spectrum, prohibit solution of the problem.

The conditions causing delay have been well established in this study and, although no clear solution is available, the subsurface conditions indicated by the delay can be recognized and the knowledge applied to engineering and groundwater problems. Within the Arcola-Tuscola area every delay which could be checked by control has coincided with the presence of sand or gravel which may be water yielding. Therefore, a thorough understanding of seismic data from unconsolidated sediments can increase the usefulness of the refraction seismograph for groundwater exploration.

The calculation of depth below the delay-causing zone has been a major problem. Brown and Robertshaw[23] have described a method using an average velocity for the unconsolidated materials, which requires an assumed velocity for the low-speed zone. It is claimed that "reasonable results" may be obtained by the method if the thickness of the low-velocity layer and that of the overlying layer are of a "similar order."

[21]Barkson, Joseph, Personal communication, December 1953.
[22]Sommers, John, Personal communication, December 1953.

[23]Brown, P. D., and Robertshaw, J., *op. cit.*, pp. 351-352.

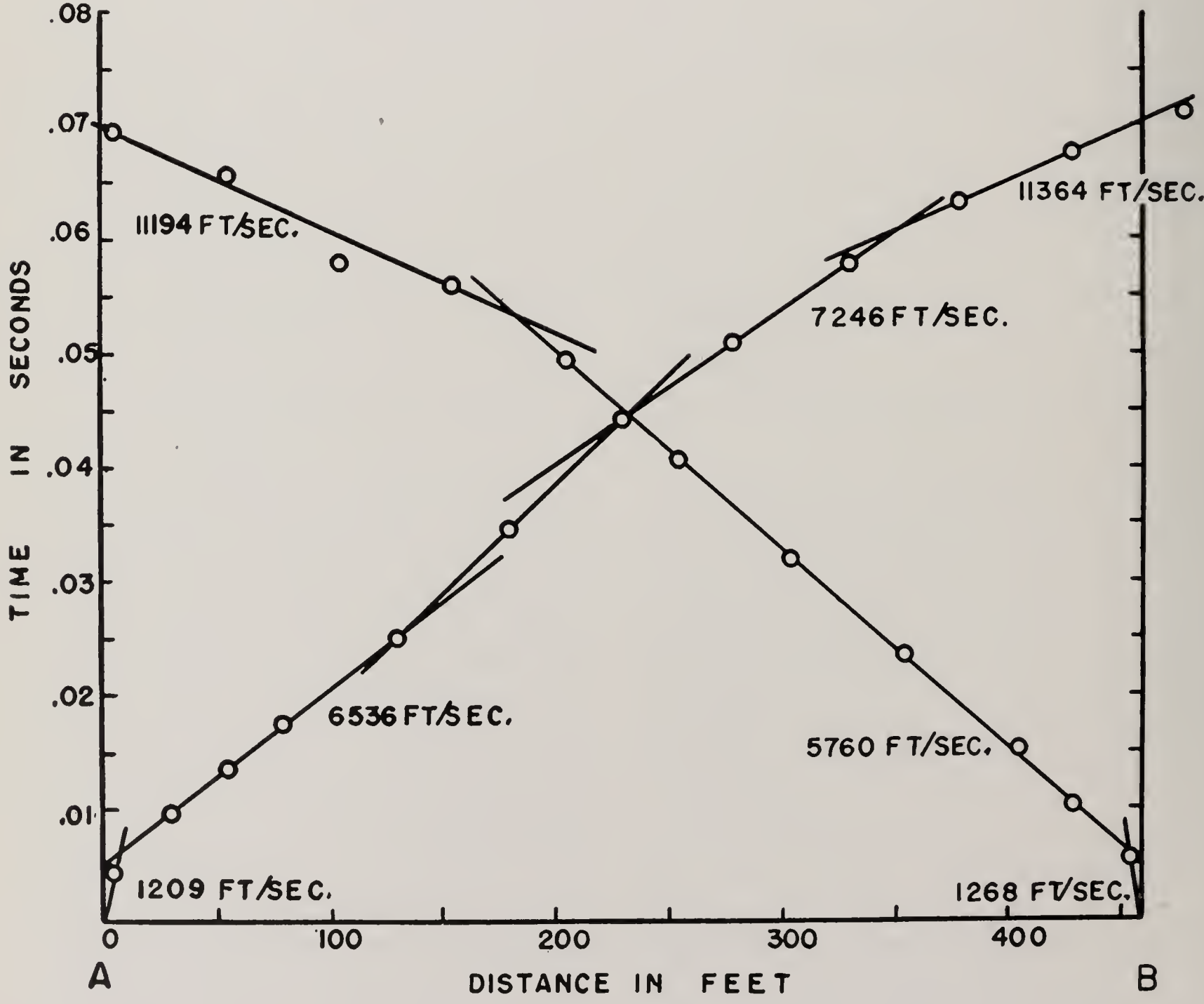


Fig. 21.—Travel-time graph for station 80, sec. 10, T. 14 N., R. 8 E., Arcola quadrangle.

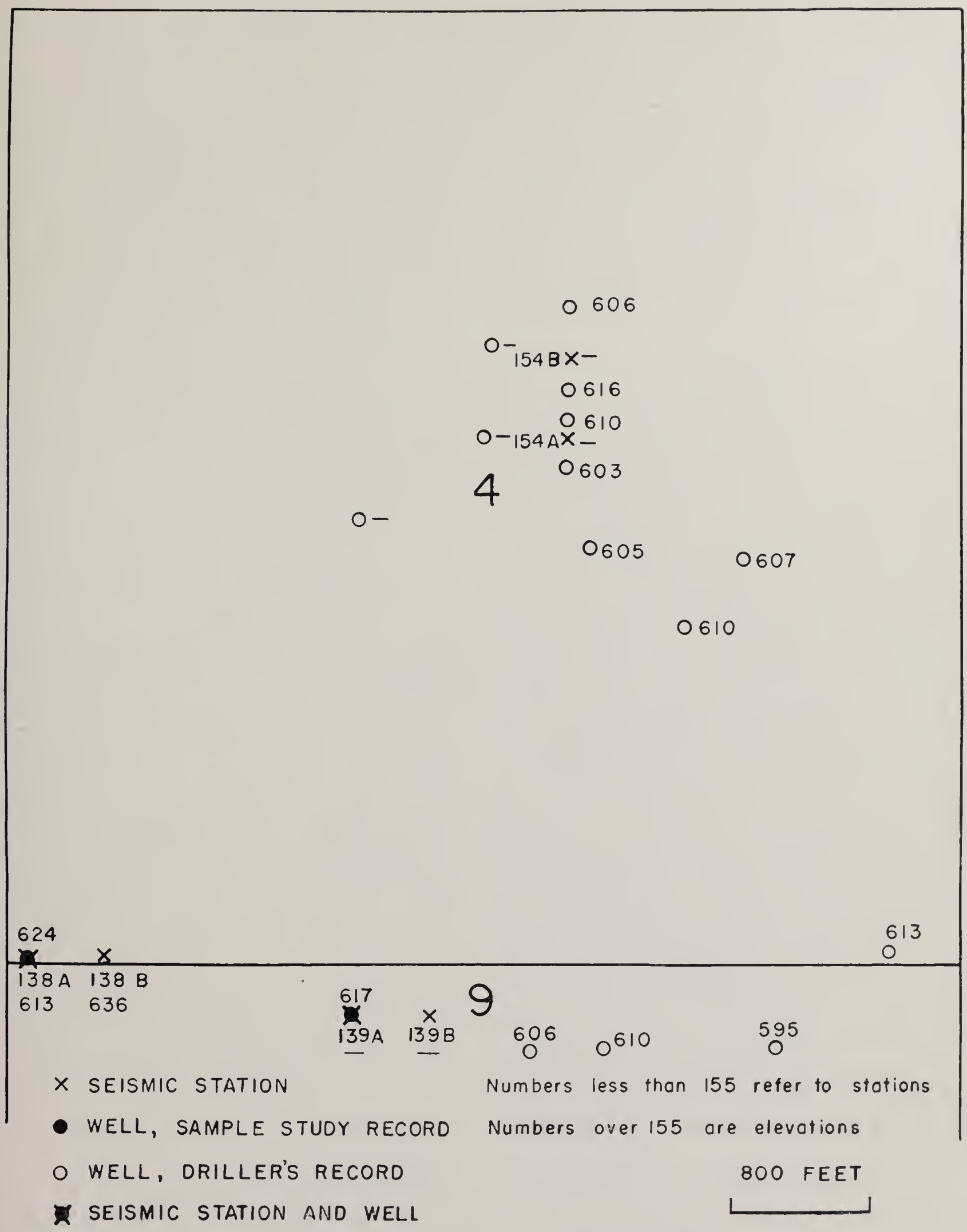


FIG. 22.—Sec. 4 and part of sec. 9, T. 14 N., R. 8. E. (*see* plate 1).

Interpretation and calculation in delay problems have been discussed by Slichter.[24] His solution demands a continuous increase of velocity in the layer overlying the low-velocity zone. Such a condition results in curved ray paths in this layer, which are represented on the travel-time graph by downward-curved velocity segments. None of the velocity segments in any of the travel-time graphs obtained in this study curve downward; all are straight-line curves.

A new empirical method of obtaining depths below the delay zone is illustrated by the wave-front diagram and travel-time curve shown in figure 19, which shows a hypothetical case in which there are four velocity layers or zones. The V_3 layer transmits seismic energy at 4000 ft./sec., 1000 ft./sec. slower than the transmission velocity in the overlying V_2 layer. According to refraction seismic theory, the buried low-velocity layer cannot have a representative line on the travel-time curve. The travel-time graph shows only the V_1, V_2, and V_4 velocities, but the V_3 layer, although not represented on the travel-time curve, does cause a change in the position of the V_4 line because of its slowing effect on the waves which reach the V_4 layer. To show this effect on the V_4 line, a wave-front diagram was constructed in which the V_2 layer extended to the V_4 layer. The V_4 line has been put in the graph with a dashed line to show its position under normal three-layer conditions. The delay which would be plotted from an actual seismic record under conditions similar to those in figure 19 would appear as a gap between the 5000 ft./sec. line and the 8000 ft./sec. line.

The standard method of computing the depth to the V_4 layer is to add the thicknesses of the V_1 and V_2 layers, calculated from the travel-time graph. The increased time intercept caused by the V_3 layer results in a total computed depth of 140 feet, which exceeds the true depth by 20 feet. Thus, the standard method gives an erroneous depth determination when there is a delay. The computed apparent thickness of the V_2 layer alone, however, is 120 feet minus the true depth to the V_4 layer.

Depth to velocity interfaces above a low-velocity zone can be computed by standard procedure. Depths to the base of the low-velocity zone equal the apparent thickness of the layer overlying the low-velocity zone, such as V_2 in figure 19. To obtain depths to velocity interfaces below the low-velocity zone, the individual thicknesses of the beds below the zone are used (see discussion of station 152 in chapter VIII).

The empirical calculation method was applied to several seismic computations involving delays at stations where there was nearby control. The A end of station 139 in sec. 9, T. 14 N., R. 8 E., was at the site of a test well (fig. 22). The travel-time curve at the station is shown in figure 20. The driller's log of the well follows:

E. C. Baker and Sons, Arcola City Test Hole 8-53, NE ¼ NW ¼ sec. 9, T. 14 N., R. 8 E., Douglas Co., elev. 670 feet (estimated from topographic map).

	Depth in feet
Soil	0–1
Yellow clay	1–12
Blue clay	12–53
Blue sandy clay	53–85
Sand and gravel	85–94
Mud sand	94–110
Sand and gravel	110–112
Clay	112–116
Limestone	116–117

The depth to bedrock calculated from seismic data by the empirical method is 117 feet, just one foot deeper than the bedrock in the well. If the 117 feet were added to the 8-foot thickness of the V_1 layer, the total depth of 125 feet is in considerably greater error than the total arrived at by the empirical method.

Seismic data from station 138 in the SW ¼ of sec. 4, T. 14 N., R. 8 E., can be compared with data from the well at 138A (fig. 22). There is no indication of a delay on the travel-time curve at the A end but a delay does appear on the B end. The log for the well, given on page 38, shows only 10 feet of pack gravel and mud sand overlying bedrock. Judging from the absence of a delay at the A end, this is not

[24]Slichter, L. B., The theory of the interpretation of seismic travel-time curves in horizontal structures: Physics, vol. 3, pp. 273-295, 1932.

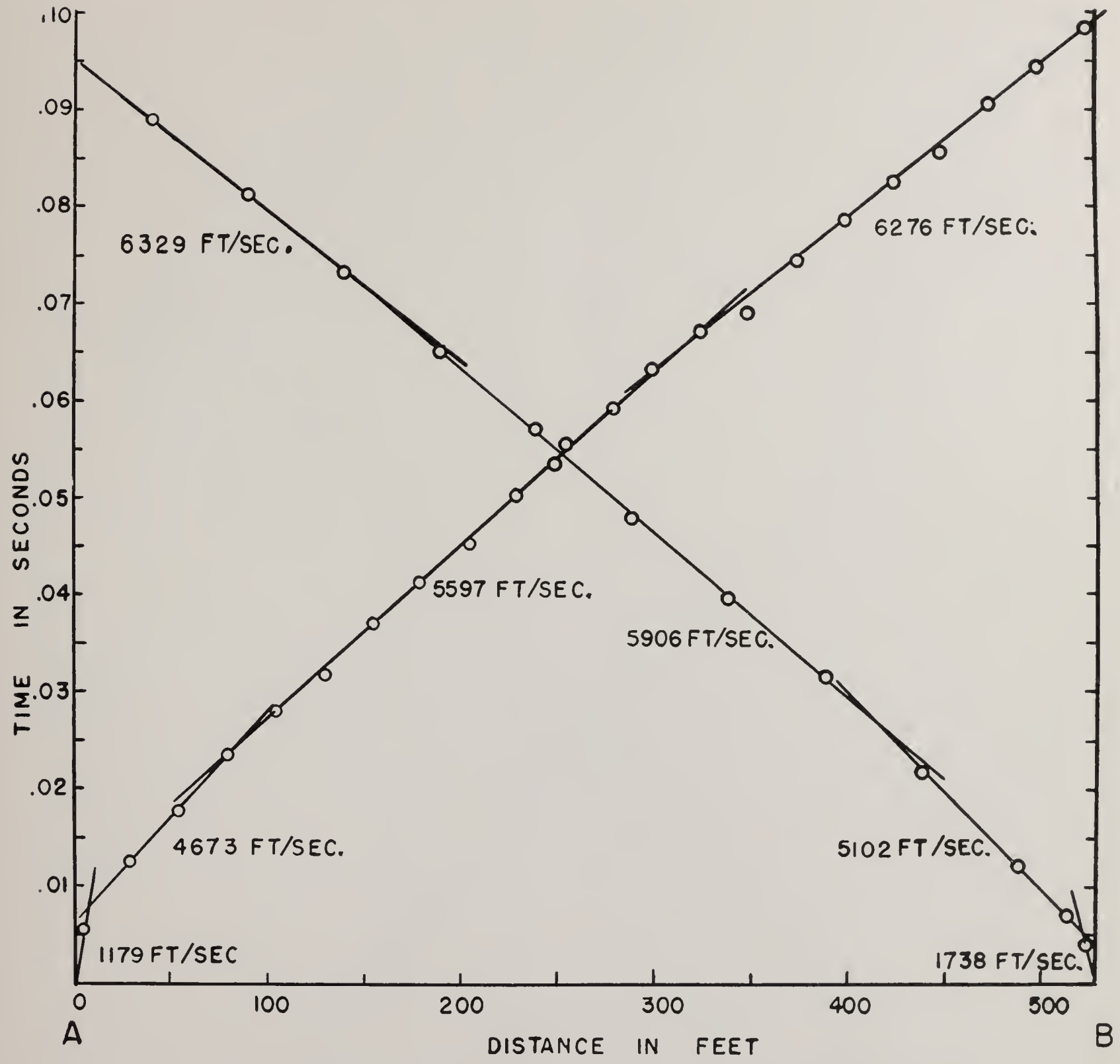


FIG. 23.—Travel-time graph for station 151, sec. 31, T. 26 N., R. 3 W., Mackinaw quadrangle.

a great enough thickness of low-velocity material to cause a delay. In the test well at station 139A, which is 1800 feet southeast of the well near station 138, 27 feet of low-velocity material was encountered. The depth to bedrock in the well at station 138 is 90 feet and in the well by station 139 it is 116 feet, indicating that the bedrock surface dips to the east from the well at station 138. The depth to bedrock at 138B would, therefore, be between 90 feet and 116 feet. The calculated depth at 138B by the empirical method is 104 feet, and by the conventional method of totaling thicknesses it is 139 feet, which is much deeper than any recorded bedrock top in the area. The delay is recorded at 138B because of the greater thickness of the sand and gravel section overlying bedrock than at 138A. The top of the sand and gravel between the two test wells appears from the well logs to be nearly horizontal, thus it is likely that there is about 20 feet of the low-velocity material at 138B.

In the preceding two examples of delay curves, the delays were between the bedrock velocity segment and the unconsolidated material velocity segment, indicating that the low-velocity material was immediately above the bedrock surface, which is substantiated by the well logs. The travel-time graph of station 80 (fig. 21), in sec. 10, T. 14 N., R. 8 E., in the Arcola quadrangle, suggests a low-velocity zone at

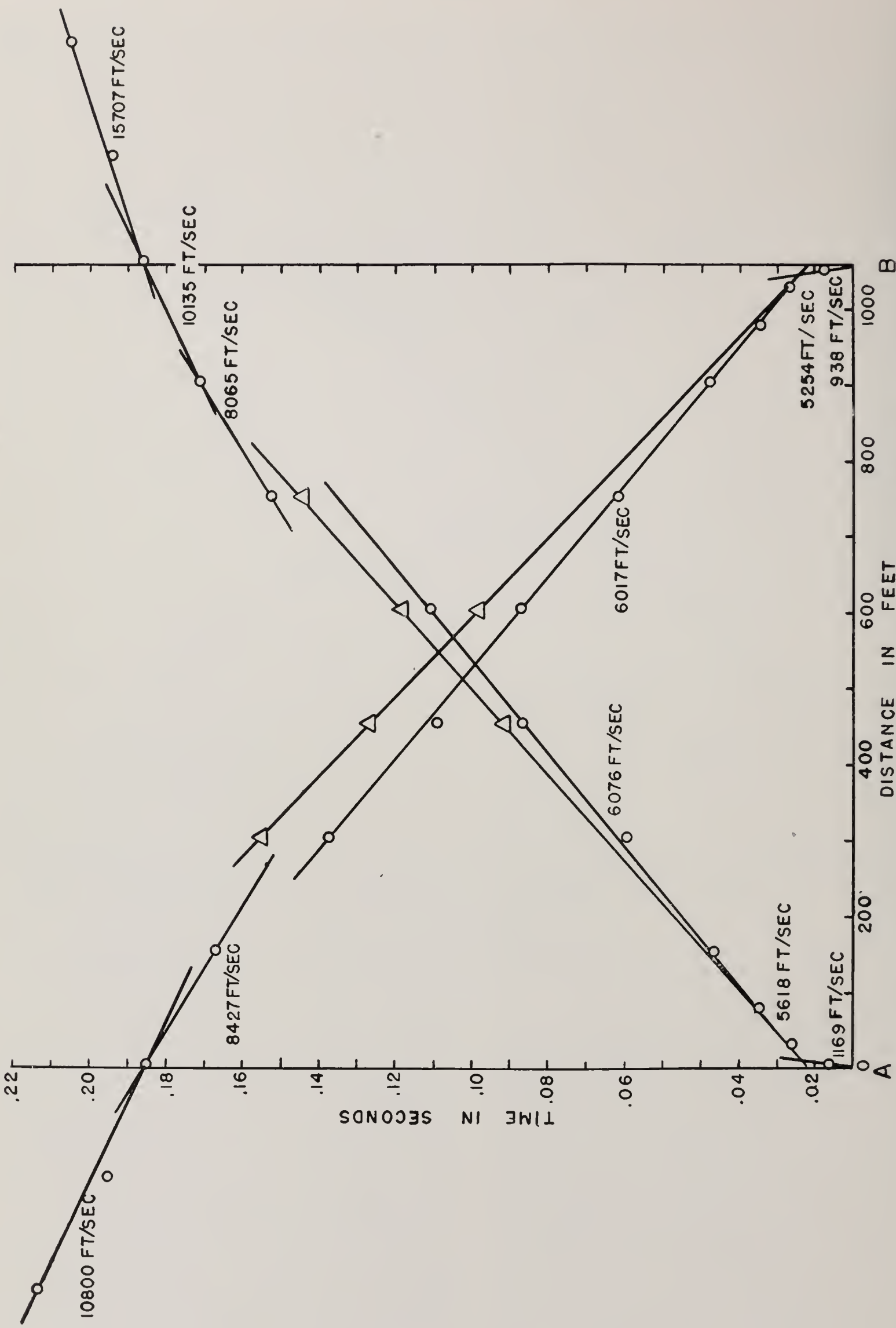


Fig. 24.—Travel-time graph for station 152, sec. 18, T. 19 N., R. 8 E., Mahomet quadrangle.

the A end between layers assumed to be of Wisconsin and Illinoian ages, rather than immediately over the bedrock. The curves for station 80 show the delay to be less abrupt than the delays at other stations. The travel time measured by the sixth geophone lies on the delay segment of the curve, which is rather unusual. Calculation of the top of the 7246 ft./sec. (Illinoian) layer gives a depth of 57 feet by the empirical method, which conforms satisfactorily to data from a nearby well. The calculated depth to bedrock is 109 feet. It might be assumed that the depth calculations are incorrect for a curve such as that for station 80B, which does not show a delay similar to that found on the opposite reversed profile, for the wave energy recorded must travel through the low-velocity material to return to the surface. This assumption, however, does not appear to be valid. A test well was drilled at the B end of station 80, where no delay was recorded. The depth to bedrock found in the well was 83.5 feet and the calculated depth was 84.6 feet.

Travel-time graphs on which only one profile exhibits a delay may denote the presence of a low-velocity layer that either pinches out or changes in lithology as the end of the curve showing no delay is approached. When both profiles have a delay, the low-velocity layer is believed to be continuous throughout the length of the profile.

Reference has been made to the shot-hole depth correction used in depth calculations. When a dynamite charge is placed below the surface of the ground, a correction must be made for the difference between the geophone level and the dynamite level. A simple geometrical correction has been used in all work done in this study. It consists of adding half the shot-hole depth to the final depth summation for any interface being measured. The time it takes a wave to travel from the dynamite in a shot hole to the geophones at the ground surface is the same as it would be if the shot depth is moved up half its distance and the geophones down the same distance, thereby placing them on the same hypothetical plane half way up the shot hole. It is then necessary to add the remaining one-half of the shot depth to raise the plane to the ground surface. This correction is commonly used in engineering geology applications[25] of the refraction seismograph.

The depth measured by refraction seismic means is not the depth directly below the shot point. It is displaced some distance in the direction in which the geophones are laid out, the amount being a function of the thicknesses of the layers and the ray-path angles at each velocity interface (figure 2). This depth displacement explains why calculated depths for two profiles sharing one shot point may be different, as at stations 16 and 17 in Tuscola quadrangle (table 2) and stations 141 and 155 in Arcola quadrangle (table 1).

Many of the causes for error in refraction seismic depth computations have been discussed. Errors may also arise from misinterpretation of valid data or from the necessity of having to use insufficient data, or both. The greater the number of seismic stations in any area, the greater the accuracy, for incorrect interpretations often become evident and can be corrected. The lack of sufficient data may be the result of widely spaced shots or poor subsurface control. Bedrock depth determinations in the Tuscola quadrangle contain higher probable error factors than those in the Arcola quadrangle because the depths are greater and there is less control. Reasonable accuracy has been attained in determining depths to velocity changes within the unconsolidated material. Bird,[26] who worked at depths of less than 100 feet, usually less than 50 feet, believes that a 10 percent error in depth is to be expected in refraction work. Coster and Gerrard[27] refer to a "standard error of ± 10%."

Seismic exploration on glacial deposits which may rapidly change laterally in lithology is liable to calculation errors when

[25]Linehan, D., *op. cit.*, pp. 167-170, 1952.
[26]Bird, P. H., Personal communication, September 1953.
[27]Coster, H. P., and Gerrard, J. A. F., A seismic investigation of the history of the River Rheidol in Cardiganshire: Geol. Mag., vol. 84, no. 6, p. 364, 1947.

shots cannot be closely spaced. Occasionally, one end of a reversed profile will have one more velocity layer represented than the other, denoting a discontinuous velocity layer. The calculation equations for a reversed profile require that the velocity layers correspond at both ends of the station. If this condition is not met, the depths must be computed by the shorter single-end form. The accuracy for the shallow, relatively flat-lying Pleistocene surfaces has not been significantly affected by such a procedure, but the deeper bedrock depths often must be accepted with caution.

CHAPTER VIII — CORRELATION OF SEISMIC AND GEOLOGIC DATA

The primary aim of this study is to evaluate the shallow refraction seismograph in the differentiation of Pleistocene deposits. The Arcola-Tuscola area in central Illinois, in Champaign, Coles, and Douglas counties (fig. 1), was selected for this investigation because it contains an average amount of subsurface control and has a level land surface ideally suited to refraction seismic work. Although depth determinations may be made from seismic work done in areas with more relief, the added computations increase the chance for error. A total of 96 seismic depth determinations have been made within the two quadrangles. Additional research was conducted in other parts of Illinois.

Seismic shot locations and wells which have good logs have been plotted on plates 1 and 2. The seismic shot locations are numbered and the data for each is given in tables 1 and 2. The photographic shot records are on file at the Illinois Geological Survey, as well as the depth calculations and velocity determinations under the station number.

Elevations above sea level of the principal seismic velocity discontinuities within the unconsolidated materials are plotted on plates 1 and 2. One elevation has been plotted at each seismic station and well location. The elevation refers to what is believed to be the upper surface of the deposits of Illinoian age. The selection of these elevations has been discussed in chapter V, and figure 9 shows the generalized contours on the buried Illinoian drift plain. Dashes indicate elevations which were unobtainable either from seismic or well data.

Tuscola Quadrangle

The seismic elevations shown on plate 2 exhibit a high degree of uniformity, considering the random placement of the shot stations. A comparison of these elevations with those from the wells scattered throughout the area shows how closely they coincide with the top of the Illinoian drift plain. Horberg[1] has stated that present-day major drainage patterns follow those which existed on the Illinoian drift plain. The seismic elevations at station 33 in sec. 2, T. 16 N., R. 7 E., and station 26 in sec. 25, T. 16 N., R. 7 E., near the Kaskaskia River, provide geophysical evidence that this is so. A few seismic elevations do not seem to correlate with any of the nearby well information, but that is to be expected with any geophysical method. The notable exceptions to the otherwise uniform surface are station 11B, sec. 22, T. 17 N., R. 8 E., station 13B, sec. 5, T. 15 N., R. 8 E., and station 95A, sec. 13, T. 16 N., R. 7 E.

On plate 2, seismic elevation data have been plotted for an additional velocity break within the Pleistocene deposits. Its elevation corresponds with a significant velocity interface between the surface of the ground and the principal contact shown on plate 2. These upper velocity interfaces are shown in the first elevation column of table 2, as they are believed to represent the contact between drifts deposited during Tazewell time by the Shelbyville ice sheet and by the later Cerro Gordo ice sheet. The uppermost velocities at stations 9 and 84 are similar to those at stations 41 and 113, and all are less than the average Illinoian velocities shown in table 3. Therefore, the elevations at stations 9 and 84 also are assigned to the Cerro Gordo-Shelbyville contact. The absence of such velocities elsewhere in the area is believed to be due to the fact that the Cerro Gordo drift in most places is less than the minimum thickness required for its appearance on the seismic record. If the intermediate velocities at stations 9, 41, 84, and 113 represent Cerro Gordo drift, then the rest of the velocities assigned to materials of Wisconsin age represent drift of Shelbyville age.

[1]Horberg, Leland, Pleistocene deposits below the Wisconsin drift in northeastern Illinois: Illinois Geol. Survey Rept. Inv. 165, pp. 36, 38, 1953.

TABLE 1.—Arcola Quadrangle Seismic Data
(Elevations in feet above sea level)

Station	*Location*	*Cerro Gordo-Shelbyville Contact*	*Wisconsin-Illinoian Contact*	*Top of Bedrock*	*Seismic Indication of Possible Aquifer*
60A	Sec. 23, T. 13 N., R. 7 E.		666	627	
60B			660	619	
69	Sec. 27, T. 13 N., R. 8 E.			604	
70	Sec. 27, T. 13 N., R. 8 E.		661		x
71	Sec. 22, T. 13 N., R. 8 E.	678	661	618	
72	Sec. 22, T. 13 N., R. 8 E.		630	524	
73	Sec. 15, T. 13 N., R. 8 E.			557	x
74	Sec. 15, T. 13 N., R. 8 E.		601	593	x
75	Sec. 10, T. 13 N., R. 8 E.			600	
76	Sec. 3, T. 13 N., R. 8 E.			605	x
77	Sec. 3, T. 14 N., R. 8 E.	651		591	
78A	Sec. 3, T. 14 N., R. 8 E.			587	
78B				594	
79	Sec. 11, T. 14 N., R. 8 E.			585	
80A	Sec. 10, T. 14 N., R. 8 E.		608	556	x
80B				580	
81	Sec. 15, T. 14 N., R. 8 E.			583	
82	Sec. 10, T. 14 N., R. 8 E.	646		572	
110A	Sec. 30, T. 14 N., R. 8 E.			609	
110B				630	
111A	Sec. 18, T. 14 N., R. 8 E.		606	521	
111B			620	528	
112A	Sec. 30, T. 15 N., R. 8 E.	659	603	494	
112B		666	606	488	
122A	Sec. 5, T. 14 N., R. 8 E.		592	519	
122B			625	509	
123A	Sec. 8, T. 14 N., R. 8 E.			604	
123B				609	
124A	Sec. 8, T. 14 N., R. 8 E.		624	548	
124B			621	586	
125A	Sec. 5, T. 14 N., R. 8 E.				
125B					
138A	Sec. 4, T. 14 N., R. 8 E.		613		
138B			636	567	x
139A	Sec. 9, T. 14 N., R. 8 E.			553	x
139B				581	x
140A	Sec. 10, T. 14 N., R. 8 E.			580	
140B				603	
141A	Sec. 33, T. 15 N., R. 8 E.			581	
141B				583	
154A	Sec. 4, T. 14 N., R. 8 E.			558	x
154B					
155A	Sec. 33, T. 15 N., R. 8 E.	657	607	541	
155B				583	

TABLE 2.—Tuscola Quadrangle Seismic Data
(Elevations in feet above sea level)

Station	*Location*	*Cerro Gordo-Shelbyville Contact*	*Wisconsin-Illinoian Contact*	*Illinoian-Kansan Contact*	*Top of Bedrock*	*Seismic Indication of Possible Aquifer*
6A	Sec. 30, T. 16 N., R. 8 E.		622		459	
6B			629		514	
7A	Sec. 21, T. 16 N., R. 8 E.		618		460	
7B			610	562	468	
8A	Sec. 17, T. 16 N., R. 8 E.		625	530	422	x
8B			616	509	382	x
9A	Sec. 8, T. 16 N., R. 8 E.	663		563	349	
9B		664		537	341	
10A	Sec. 36, T. 17 N., R. 7 E.				546	
10B					531	
11A	Sec. 22, T. 17 N., R. 8 E.				555	
11B		666			558	
12A	Sec. 28, T. 17 N., R. 8 E.		632		507	
12B			645		581	
13A	Sec. 5, T. 15 N., R. 8 E.		632		433	
13B			651		319	
16A	Sec. 33, T. 16 N., R. 8 E.		627		463	
16B			621		508	
17A	Sec. 33, T. 16 N., R. 8 E.		620		486	
17B			621		583	
22A	Sec. 3, T. 16 N., R. 7 E.		653		543	
22B			645		434	
23A	Sec. 30, T. 16 N., R. 8 E.		608		342	
23B			619		385	
26A	Sec. 25, T. 16 N., R. 7 E.		608		544	
26B			615		480	
33A	Sec. 2, T. 16 N., R. 7 E.		622		531	
33B					588	
40A	Sec. 31, T. 17 N., R. 8 E.				503	
40B					458	
41A	Sec. 19, T. 17 N., R. 8 E.	659	633		419	
41B		654	643		431	
42A	Sec. 29, T. 17 N., R. 8 E.				400	x
42B					379	x
43	Sec. 3, T. 15 N., R. 8 E.		623		523	
44A	Sec. 20, T. 17 N., R. 8 E.		632		244	
44B			632		279	
45A	Sec. 27, T. 17 N., R. 7 E.		640		450	
45B			641		414	
46A	Sec. 22, T. 17 N., R. 7 E.		636		411	
46B			639		452	
47A	Sec. 32, T. 17 N., R. 7 E.		645		502	
47B			645		492	
48A	Sec. 29, T. 17 N., R. 7 E.		646		350	
48B			638		386	x
84A	Sec. 5, T. 16 N., R. 8 E.	667			435	
84B		675			369	
95A	Sec. 13, T. 16 N., R. 7 E.		583		435	
95B					532	
96A	Sec. 15, T. 16 N., R. 7 E.		643		540	
96B					582	
113A	Sec. 18, T. 15 N., R. 8 E.	668	600		517	
113B		662	628		509	

TABLE 3.—Velocities in Arcola and Tuscola Quadrangles

	Aerated	*Cerro Gordo*	*Shelbyville*	*Illinoian*
Tuscola	1165 ft./sec.	4963 ft./sec.	5834 ft./sec.	7203 ft./sec.
Arcola	1379 ft./sec.	4761 ft./sec.	5852 ft./sec.	7187 ft./sec.

The velocities in the aerated or oxidized zone, in the presumed Cerro Gordo drift, in the presumed Shelbyville drift, and in the Illinoian drift, in the Tuscola quadrangle, have been totaled and averaged. Fifty-three aerated zone velocities average 1165 ft./sec.; eight velocities of the presumed Cerro Gordo deposits average 4963 ft./sec.; forty-nine presumed Shelbyville velocities average 5834 ft./sec.; and forty-three Illinoian velocities average 7203 ft./sec. Comparison shows close correlation of these velocities with those in the Arcola quadrangle (table 3).

(For a discussion of the Arcola velocity averages see chapter VII.) The comparison in table 3 emphasizes the uniform velocity differentials that permit the mapping of Pleistocene stratigraphic breaks with the refraction seismograph.

No velocities representative of deposits of Illinoian age are present on records from stations 9, 10, 11A, 33B, 40, 42, 84, 95B, and 96B. These stations extend from station 11 in sec. 22, T. 17 N., R. 8 E., in the northeast, to station 96 in sec. 15, T. 16 N., R. 7 E., in the southwest, in a pattern suggestive of a drainage system. It is not known if the absence of the velocities indicates a buried channel in the Illinoian drift because there is no available control in this area, but it is probable.

A group of five elevations were recorded, at stations 7B, 8A and B, and 9A and B in secs. 21, 17, and 8 of T. 16 N., R. 8 E., which do not correlate with other surfaces described. They are shown in the column headed Illinoian-Kansan contact in table 2. They could be bedrock-top elevations, but the deeper elevations in the adjoining column correspond better to the bedrock drainage pattern outlined by other seismic stations and wells. Kansan tills have been described in some of the deeper wells in the area, and it is possible that the five elevations may represent the surface deposits.

The elevations of the base of the aerated zone coincide with or are slightly higher than the elevations of the base of the oxidized zones reported by drillers in nearby wells. It is quite possible that the surface formed by these elevations may be the top of the zone of saturation; water was often encountered in auger borings for shot holes. Lester[2] has found this to be the case at times, and in such instances, his definition of the aerated zone is the same as that by Meinzer.[3] However, since Meinzer's aerated zone is measured from the water table and Lester's zone is a seismic velocity change, the two are not the same when the water table lies, possibly temporarily, below the elevation of seismic velocity change. The seismic aerated zone therefore corresponds with the surface zone of oxidation. The elevations between this oxidation zone and the zone of saturation may vary from place to place.

Arcola Quadrangle

The notations used for seismic shot and well data on plate 1 for the Arcola quadrangle are the same as those used on plate 2 for the Tuscola quadrangle. A detail map of section 4 and part of section 9, T. 14 N., R. 8 E., shows the closely spaced wells in that area (fig. 22).

The majority of the seismic locations do not indicate the presence of Illinoian drift velocities, probably because of the shallow depth to bedrock throughout most of the area. Whenever present, seismic elevations representative of the top of the Illinoian drift correspond reasonably well with the surface shown in well logs, although the

[2]Lester, O. C., Jr., Seismic weathered or aerated layer: Am. Assoc. Petr. Geol. Bull., vol. 16, no. 12, p. 1234, 1932.
[3]Meinzer, O. E., The occurrence of ground water in the United States: U. S. Geol. Survey Water Supply Paper 489, pp. 76-82, 1923.

exceptions are more apparent because of the smaller number of Illinoian surface elevations. Stations 72, 80A, 111A, 112, 122B, 124, 138A, and 155A show the best correspondence. Two elevations have been calculated for the Illinoian top near the well in the SW cor. sec. 4, T. 14 N., R. 8 E., the log of which is given on page 38.

The seismograph recorded the top of the blue sand clay at the 58-foot depth at station 138A, but at station 122B and station 124AB the top of the hardpan zone at the 47-foot depth was recorded. The hardpan does not appear in nearby wells to the east, but a southwestern extension of it may account for the corresponding elevation found at station 111B in section 18 of the same township. Station 111A records the general Illinoian surface elevation of the area.

Seismic velocities lower than the average Wisconsin or Illinoian velocities have been noted at twelve shot locations within the Arcola quadrangle. The occurrence of thin layers of this low-velocity material over the higher velocity Wisconsin drift in several stations is evidence for the previously stated assumption that the lower velocities represent Cerro Gordo age drift. Additional evidence was found at station 71, T. 13 N., R. 8 E., just north of the crest of the Cerro Cordo moraine. The elevation of the Shelbyville-Cerro Gordo contact is 678 feet, computed by the seismic method. The undissected surface of the Shelbyville drift south of the moraine in this area ranges from 670 to 680 feet approximately. The next higher average velocity is believed to be representative of Shelbyville age drift.

The elevation of the Illinoian surface in the vicinity of Arcola ranges between 600 and 615 feet. Well logs from the northwestern part of the quadrangle near the town of Arthur and well logs and seismic data from the Tuscola quadrangle to the north (plate 2) show a rise in the elevation of the Illinoian to about 645 feet. The northward rise accompanies a gradual rise in the land surface.

Seismic Research Outside the Arcola-Tuscola Area

Ideally suited for checking the refraction seismic method is the area at the Farmdale railroad cut in the SW ¼ NW ¼ sec. 31, T. 26 N., R. 3 W., in the Mackinaw quadrangle, Tazewell County. There is a relatively simple, well-exposed stratigraphy, a uniform lateral extent of the stratigraphic units, a level stretch of land close to the exposure for making the seismic profile, and a reliable geologic description of the exposure. Pleistocene exposures in the area have been studied since the latter part of the nineteenth century. Leverett[4] illustrated and discussed an exposure later well known as the "Farm Creek exposure" in the SE ¼ sec. 30, T. 26 N., R. 3 W., of the same quadrangle. The exposure was later described in detail by Leighton[5] who considered it to be remarkable because it shows evidence of the Wisconsin and Illinoian glacial epochs and the Sangamon and Peorian interglacial epochs. The Farmdale railroad cut, recently excavated for the Toledo, Peoria and Western Railroad, has exposed the same sequence. The stratigraphic sequence has been carefully examined and described by Leighton and Willman[6] following the modern classification shown in figure 8. The section on the north side of the cut near the west end measured downward from an elevation of 640 feet is shown at the top of p. 56.

Seismic station 151 was along the south side of the east-west road on the undissected portion of the plateau in the N½ SW ¼ sec. 31, T. 26 N., R. 3 W., parallel to and 250 feet from the southern rim of the railroad cut. A reversed profile was made with a distance of 530 feet between shot points. The thicknesses of three velocity layers were calculated for each end. At the west end, depths to velocity interfaces are 7.2 feet, 20.3 feet, and 51.3 feet. At

[4]Leverett, Frank, The Illinois glacial lobe: U. S. Geol. Survey Mon. 38, pp. 128, 187, 1899.

[5]Leighton, M. M., A notable type Pleistocene section, the Farm Creek exposure near Peoria, Illinois: Jour. Geol., vol. 34, pp. 173-174, 1926; reprinted as Illinois Geol. Survey Rept. Inv. 11, 1926.

[6]Leighton, M. M., and Willman, H. B., Basis of subdivisions of Wisconsin glacial stage in northeastern Illinois: Guide Book for Field Conference by Illinois Geological Survey and Indiana Geological Survey, p. 19, 1953.

	Thickness		Total	
	Feet	*Inches*	*Feet*	*Inches*
Tazewell loess	3	6	3	6
Bloomington silt	3	6	7	
" till	8		15	
Shelbyville till	17		32	
Iowan loess	2–5		34–37	
Farmdale loess	8		42–45	
Illinoian				
horizon 1–humus zones		6–8	42–45	6
" 2–mesotil	3		45–48	6
" 3–zone of iron concentration	2	6	48–51	
" 4–till, oxidized	2–3		50–54	
" 5–till, calcareous	1–5		51–59	
Kansan (?) till	10–15		61–74	

the east end, the depths are 7.5 feet, 25.7 feet, and 54.4 feet. In the measured section, the only thickness of any magnitude between the tops of two tills appears between the top of the Shelbyville till and horizon 2 of the Illinoian sequence. This thickness ranges from 27.5 to 30.5 feet. The interval between the two deepest velocity interfaces at the west end is 31.0 feet, and at the east end it is 28.7 feet. These thicknesses correspond closely with those measured; thus the 20.3-foot and 25.7-foot interface depths are believed to represent the top of the Shelbyville till, and the 51.3-foot and 54.4-foot depths the top of the Illinoian till. Figure 23 shows the travel-time graph for station 151.

There is excellent subsurface control in secs. 18 and 19, T. 19 N., R. 8 E., in the Mahomet quadrangle, which bounds the north edge of the Tuscola quadrangle. A program of test drilling was conducted by the National Petro-Chemicals Corp. in these sections during 1951-1952, and reliable samples were received by the Illinois Geological Survey from six of the test wells. The samples, which were studied and correlated by J. W. Foster of the Illinois Geological Survey, show excellent lateral continuity of the Pleistocene stratigraphic units. The uniform stratigraphy, thick sand and gravel sections, and the overall thickness of the unconsolidated sediments were ideal for a check of the seismic method. Seismic station 152 was made near test hole number 16 in the SW ¼ SE ¼ sec. 18, with end A 30 feet north of the test well (1350 feet north of the south line of the section) and end B 1050 feet south of end A. The primary purpose of this shot was to determine the ability of the refraction seismograph to record accurately the deeper Kansan till and bedrock surfaces.

The following sample-study log shows that the Wisconsin-Illinoian contact at 95 feet should be recorded seismically along with a probable delay on the travel-time curve because of the sand and gravel from 165 to 220 feet.

Hayes and Sims, National Petro-Chemical T.H. 16, SW ¼ SE ¼ sec. 18, T. 19 N., R. 8 E., Champaign Co., elev. 703 feet (estimated from topographic map).

	Depth in feet
Pleistocene series	
Wisconsin drift	
Till, yellow to gray, silty, calcareous	0–25
Sand, yellow, medium, sorted, clean	25-30
Till, pinkish, decreasing sand below, calcareous	30–95
Illinoian drift	
Till, gray, generally very sandy, gravelly, calcareous with sand streaks	95–165
Gravel, yellow, coarse, mostly clean	165–170
Sand and gravel, yellow with clay streaks	170–200
Yarmouth sand	
Sand, yellow, medium, sorted, gravel below with soil fragments	200–220
Kansan drift	
Till, buff, very fine gravelly, calcareous	220–250
Gravel, yellow, coarse, fine-to-coarse, mostly clean	250–278
Silt, yellow, noncalcareous, wood fragments	278

The travel-time graph for station 152 is shown in figure 24. The delay caused by the approximately 55 feet of sand and gravel above the Kansan till appears on each of the reversed profiles. The expected change

of velocity at the Wisconsin-Illinoian contact does not appear on the curves. Foster[7] has stated that the Illinoian till samples examined in this area are very sandy. The high sand content has probably compensated for the increased compaction that followed deposition with the result that the Illinoian seismic velocity is similar to that of the overlying Wisconsin age drift.

Depths to the seismic velocity interfaces at each end of the station were calculated using the methods presented in this report for analyzing delay curves. The following depths were obtained.

A end	*B end*
11.1 feet	10.0 feet
17.8 feet	12.1 feet
217.0 feet	206.0 feet
288.0 feet	320.0 feet

The upper layer is the aerated zone, which is thicker than usual because artificially dredged material from the nearby Kaskaskia River has been dumped on it. The next lower surface does not correlate with changes in till noted in any of the wells in the area, although the drift overlying it has an average seismic velocity similar to that of the assumed Cerro Gordo age deposits to the south. The average velocity of 6048 ft./sec. is similar to that for Wisconsin (Shelbyville) drift in the Arcola-Tuscola area. Foster[8] has tentatively assigned the drift from 30 to 95 feet in the test hole to Shelbyville age. The next velocity change occurs at 217 feet and 206 feet, and underlying this interface is a material having an average velocity of 8246 ft./sec., which ordinarily might be considered to be bedrock. However, the well log shows the necessity of having subsurface control when doing refraction seismic work, because this depth correlates remarkably well with the top of the Kansan till described in the log. The depths of 288 feet and 320 feet to a bedrock having an average velocity of 10468 ft./sec. are less in error than was expected under the conditions. The test well was believed to have reached bedrock at a depth of 278 feet, although no sample was taken. Three adjacent wells entered Pennsylvanian shale at similar depths.

Figure 24 is an example of the use of later arrivals which occasionally are found on seismic shot records when first arrival energy is of moderate or weak strength. The points surrounded by triangles substantiate the existence of the 5618 ft./sec. and 5254 ft./sec. velocities at the A end and B end, respectively, which were not clearly indicated by first arrival data.

[7]Foster, J. W., Personal communication, December 1953.

[8]Sample study, Illinois Geol. Survey files.

CHAPTER IX — SEISMIC BEDROCK DEPTH DETERMINATION

Seismic depth-to-bedrock determinations were made at all stations used for this report. These determinations were made for two reasons: interpretative value in Pleistocene research, and bedrock information for Illinois Geological Survey files.

The bedrock elevations from all available well records are on file at the Illinois Geological Survey. The control is unsatisfactory except in the vicinity of Arcola. Seismic results range from accurate at such places as stations 79, 80B, and 81 in secs. 11, 10, and 15 of T. 14 N., R. 8 E. (plate 1 and table 1), to obviously erroneous depths at station 44 in sec. 20, T. 17 N., R. 8 E., station 13 in sec. 5, T. 15 N., R. 8 E., and station 23 in sec. 30, T. 16 N., R. 8 E. (plate 2 and table 2). Judging from the accuracy shown in bedrock determinations wherever control has been obtainable, it can be assumed that the majority of the seismic elevations err by not more than 10-15 percent, depending upon depth. Bedrock velocities were not obtained at several stations in the Arcola quadrangle, probably because of insufficient length of the geophone spread or the existence of conditions discussed in chapter VII.

That the majority of the depths are accurate is substantiated by bedrock data gathered in Madison and St. Clair counties, Illinois. A total of 78 depth determinations were made in that area to depths of 100 feet and over. The comparison of seismic data with well data has shown that errors have averaged much less than 10 percent.

The bedrock topography of the Arcola-Tuscola area was mapped and described by Horberg[1] in his *Bedrock Topography of Illinois*. Many of the control points used in this report, especially around Arcola, were not available to him when the map was being made. Horberg believed that "except for a few poorly controlled areas the interpretations of major valleys and uplands are secure." He added that "future modifications of the map will probably be largely revisions and additions of tributary valleys and secondary uplands."

Depths to the top of bedrock calculated from seismic records are shown on tables 1 and 2. Seismic data substantiate the presence of a major valley in the approximate location of Horberg's Pesotum Valley. The additional control provided by the seismic elevations indicates the presence of a previously unmapped tributary of moderate size in T. 17 N., R. 7 E. Seismic information and recent well data have resulted in the relocation of the narrow tributary that extends southward at the southeast corner of T. 16 N., R. 7 E., one mile east of the location shown on Horberg's map. The elevations of stations 113 and 112 in secs. 18 and 30 respectively of T. 15 N., R. 8 E. (tables 1 and 2) indicate the southern extension of this tributary. The seismic elevations on the Arcola quadrangle do not cover a broad enough area to permit comparison with the published map. The elevations at station 111 in sec. 18, T. 14 N., R. 8 E. (table 1), necessitate the elimination of the southward extension of Horberg's 550-foot contour.

[1]Horberg, Leland, Bedrock topography of Illinois: Illinois Geol. Survey Bull. 73, 1950.

CHAPTER X — SUMMARY AND CONCLUSIONS

The primary objective of this study has been to determine the usefulness of the refraction seismograph in recording significant horizons within Pleistocene deposits of glacial or interglacial origin. Standard field and laboratory techniques were employed except for the interpretation of abnormal seismic records. Secondary objectives have been to obtain data on depth to bedrock and to determine the usefulness of the refraction method in groundwater exploration.

Seismic depth determinations of Pleistocene deposits in the Arcola and Tuscola quadrangles conform well with at least one horizon, the prominent stratigraphic break between Wisconsin and Illinoian drifts. Examination of seismic velocities shows that the velocities representative of deposits of different ages have distinctive ranges and therefore characteristic average velocities. At a few seismic shot locations, depths were obtained which are shallower than those for the Wisconsin-Illinoian contact. Energy from the dynamite shots traveled through the material overlying the shallow interface at a consistently lower velocity than it did through the material below. Although sample study records made from well samples from the Arcola-Tuscola area do not show any differentiation between the Shelbyville or Cerro Gordo age deposits within the Wisconsin drift, the seismic depths and average velocities strongly suggest that the shallow seismic depths represent the surface between the Shelbyville deposits and thin overlying Cerro Gordo deposits. In the thicker drift sections of the Tuscola quadrangle, velocities and depths were calculated which have been assumed to show the presence of Kansan age material.

The coincidence of horizons between drifts with different average velocities and known Pleistocene stratigraphic breaks is accepted as proof of the ability of the refraction seismograph to record significant stratigraphic units within glacial drift. The success of the seismograph in such a study is dependent upon adequate subsurface control from which representative velocities for the various drift units may be correlated. An understanding of the limitations of the seismograph and the interpretive techniques are of equal importance.

Many of the seismic travel-time graphs exhibit discontinuities or delays. Evidence gathered from well logs and samples has shown the delays to be attributable to layers of sand and gravel which have lower representative seismic velocities than the overlying drift. Refraction seismic theory does not provide an explanation for the appearance of a delay from such a layer, but geologists engaged in shallow seismic exploration have encountered delays in travel-time graphs. While many recognize the presence of a low-velocity layer representing sand or gravel at some depth, others regard the delay as the result of either a buried channel or instrument failure.

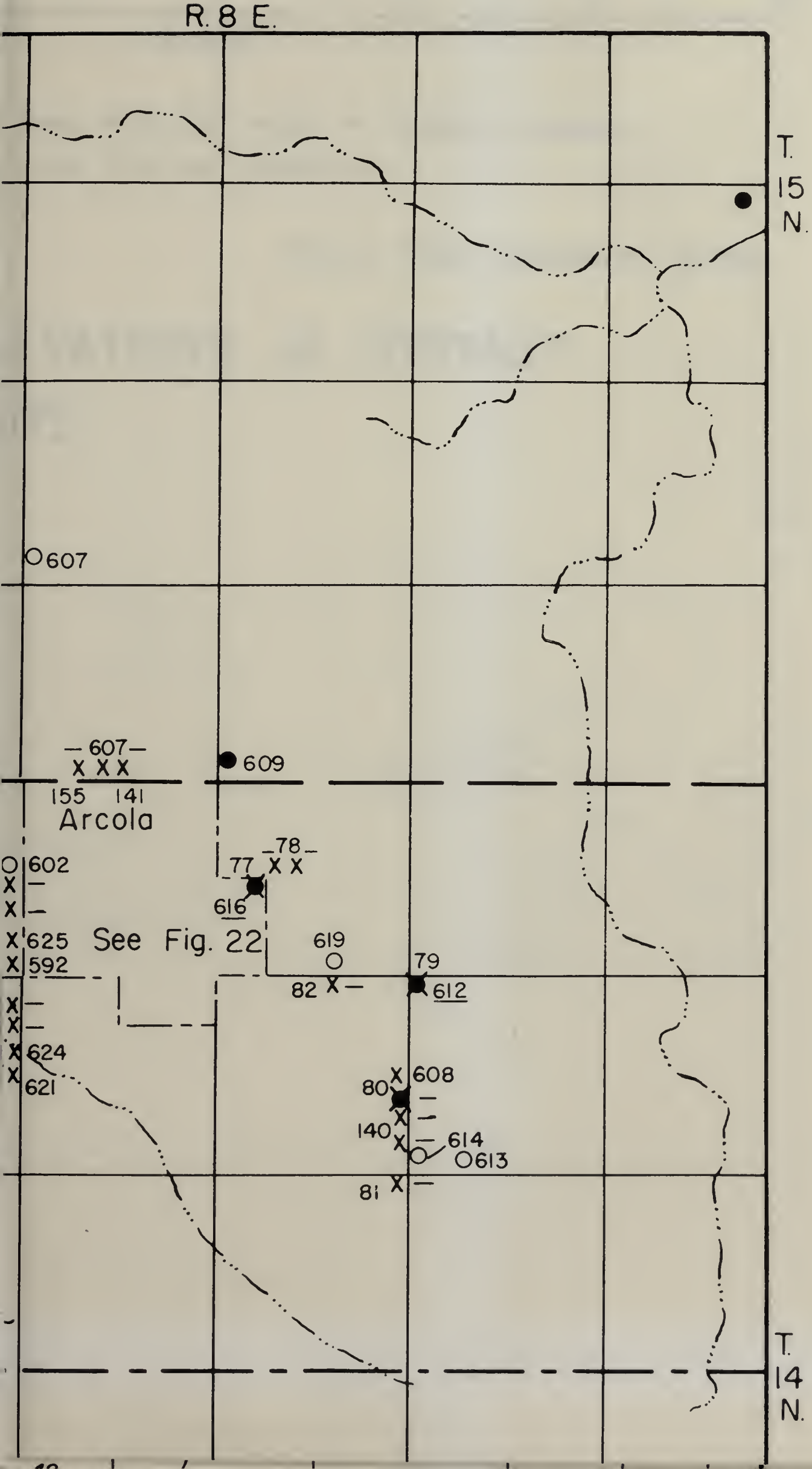
An empirical method of working out depths to velocity interfaces below the low-velocity zone was developed during this investigation as a result of study of wave-front diagrams. By using this method, the depth to the base of the low-velocity layer is obtained. Recognition of a delay in a refraction seismic record and the new method of calculating the depth to the base of low-velocity material causing the delay offers a new type of exploration to the groundwater geologist. This interpretation of travel-time curve delays applies only to work done in unconsolidated sediments.

Depths to bedrock have been obtained concurrently with depths to Pleistocene horizons for use in the development of the groundwater resources of the Arcola-Tuscola area. Many of the depth determinations have proved useful in interpreting geologic conditions in the overlying drift.

Illinois State Geological Survey
Report of Investigations No. 176
1954

Plate 1

ARCOLA QUADRANGLE



ARCOLA QUADRANGLE

x Seismic station
● Well - sample study record
○ Well - drillers record
▣ Seismic station and well

Scale of Miles
0 1 2 3

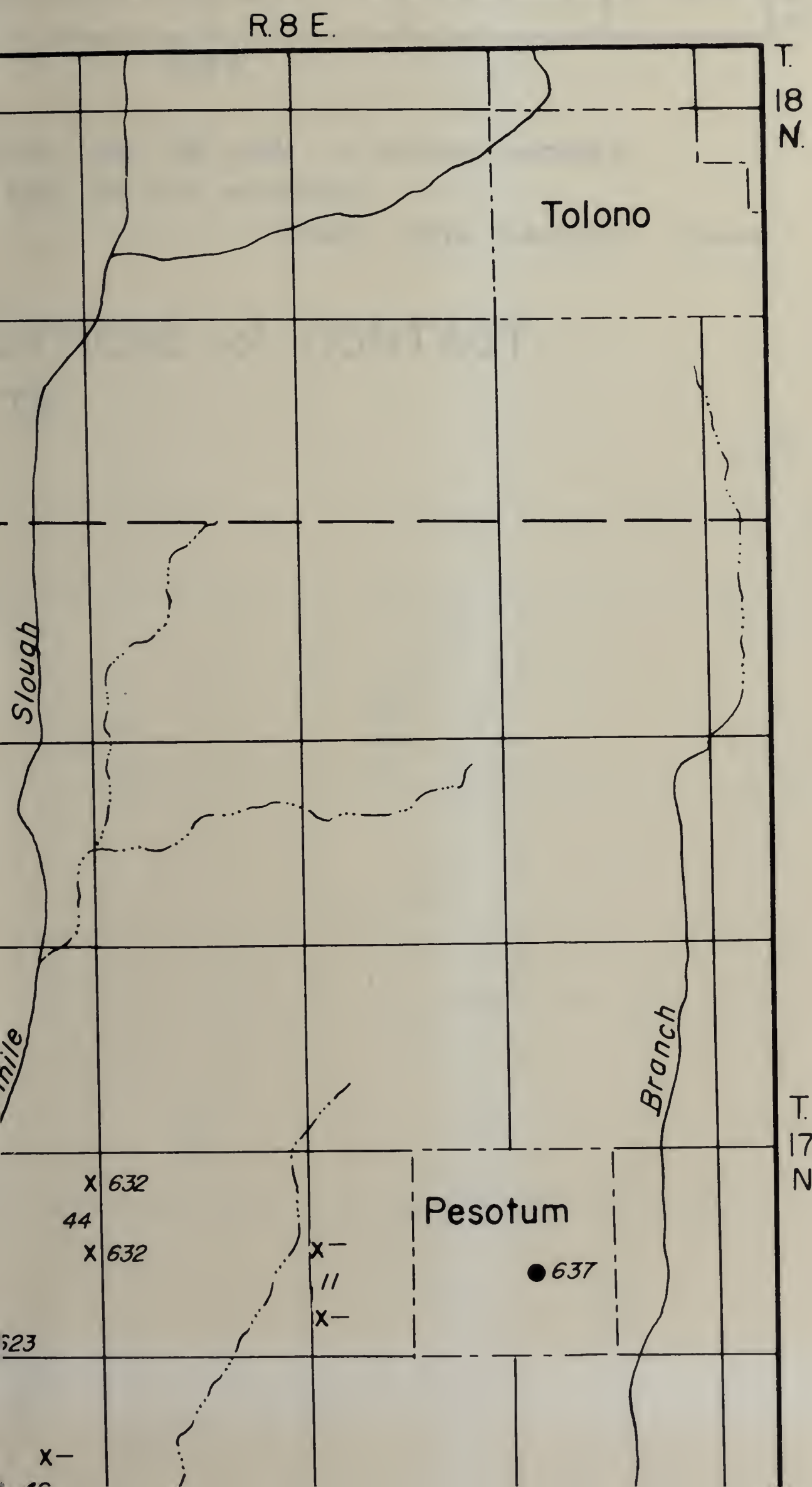
Numbers less than 155 refer to station numbers.
Numbers over 155 are elevations.

Illinois State Geological Survey

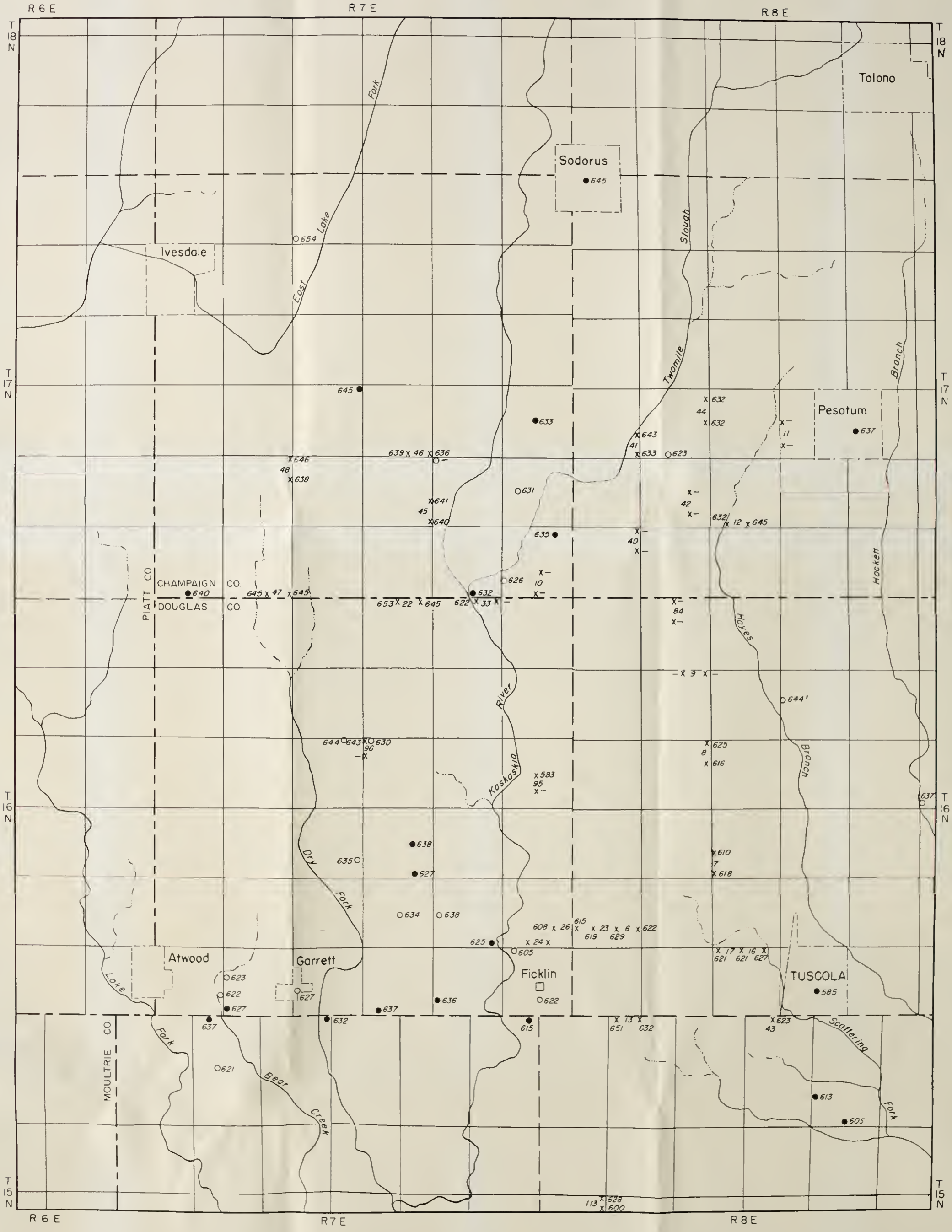
MAP SHOWING SEISMIC STATIONS and WELL LOCATIONS with ELEVATIONS of CONTACT between WISCONSIN and ILLINOIAN AGE DEPOSITS

Plate 2

TUSCOLA QUADRANGLE



TUSCOLA QUADRANGLE



x Seismic station
● Well - sample study record
○ Well - drillers record

Scale of Miles
0 1 2 3

Numbers less than 155 refer to station numbers.
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Illinois State Geological Survey

MAP SHOWING SEISMIC STATIONS and WELL LOCATIONS with ELEVATIONS of CONTACT between WISCONSIN and ILLINOIAN AGE DEPOSITS

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